COMMISSION OF THE EUROPEAN COMMUNITIES

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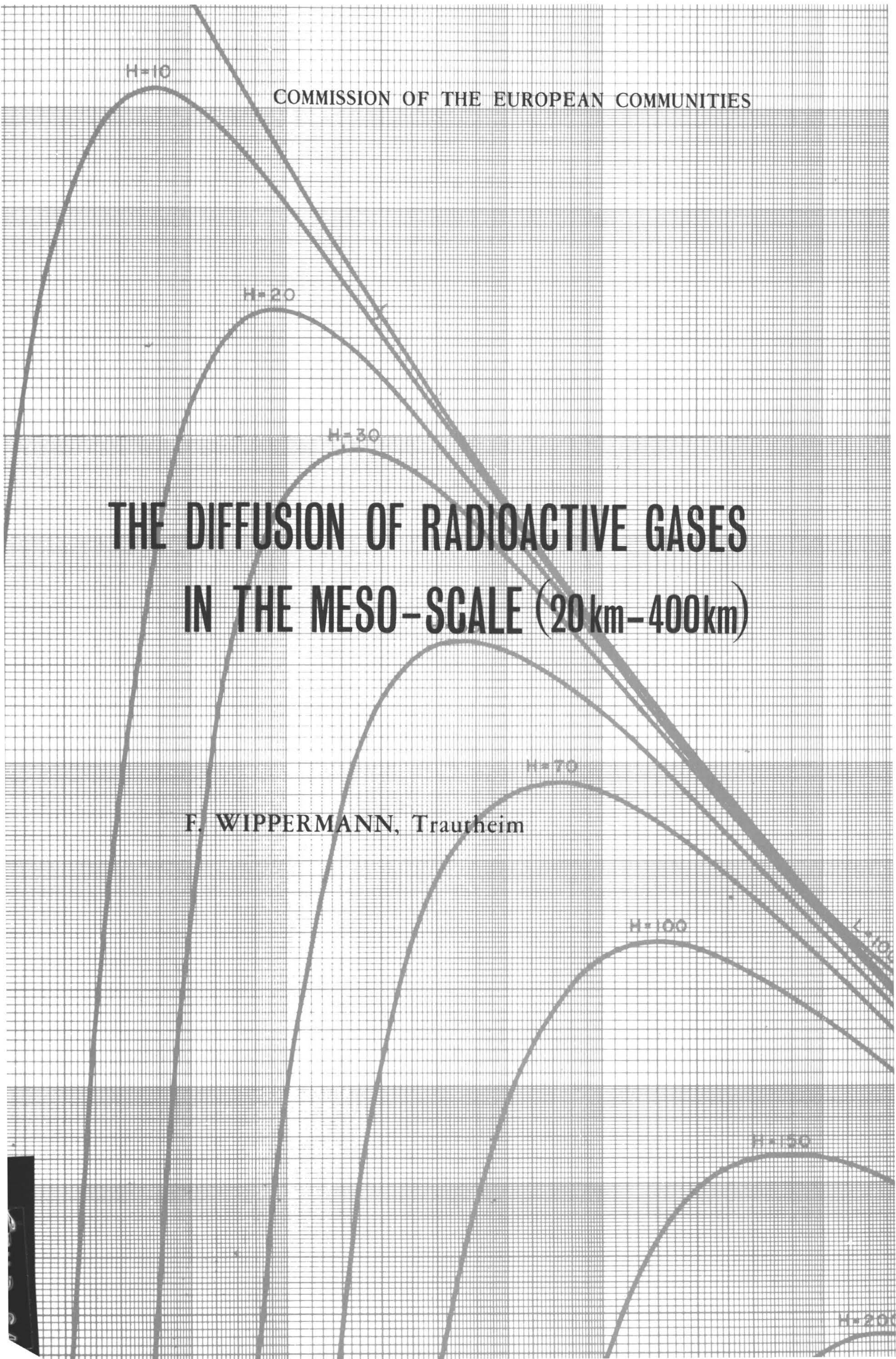
F. WIPPERMANN, Trautheim



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REPORT

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Preface

Considerable difficulties are encountered with predictions in the meteorological field (as we well know from weather forecasts), particularly when long-range forecasts (from the point of view of time or distance) are to be made. We are faced with similar problems in nuclear technology when reliable estimates of the diffusion of radioactive discharges in the atmosphere, and of any resulting radiological effects, are required.

Questions of this type also influence the opinions delivered by the Commission pursuant to Article 37 of the Euratom Treaty as to whether the intended disposal of radioactive waste from nuclear installations is likely to lead to the radioactive contamination of another Member State, and if so, to what extent. In view of the increasing role played by nuclear power in energy programmes, and thanks to a heightened consciousness of environmental matters, calculations of long-range diffusion in the atmosphere have recently aroused great interest; firstly because the possible overlapping of effects produced by separate emission sources can no longer be ignored, and secondly because there is an increasing need to determine more accurately the contamination of the population at large, and not merely that of the population group living near the source of emission.

The efforts which have been made for some years by the Directorate for Health Protection in collating information to further our understanding of the complex processes involved in the atmospheric transfer of radioactive matter are to be seen in the light of these problems. It is hoped that this study may indicate how progress may be made in the quantitative analysis of rarefied gaseous discharges transported over large distances, i.e. several hundred kilometres, and hence provide an improved basis for estimates of contamination.

Dr. P. RECHT

S u m m a r y

The term "Mesoscale" refers to distances between 20 km and 400 km from the source; in defining this range, the structure of atmospheric turbulence is taken into account.

To arrive at an evaluation of diffusion in the mesoscale, quantitative methods from the microscale (source distance $<$20 km) and qualitative methods from the macroscale (source distance $>$400 km) are extrapolated into the mesoscale. In the first case a table is given to read off the minimum factor by which the concentration is reduced in the mesoscale as the source distance increases; to obtain the diffusion for the worst possible case, the existence of a mixing-layer topped by a temperature inversion, was assumed. For this it was essential, first of all, to determine the source distance x_D, beyond which the diffusing gases are completely mixed within the mixing-layer of thickness D. To make allowance for all possible thicknesses of this mixing-layer, a measurement carried out at ground level at only 10 km from the source can be used to calculate the correct concentrations in the mixing-layer; the dilution factors will then be related to this value.

Possible ways of an improved incorporation of certain factors in the diffusion estimate, such as the topography of the earth's surface, the roughness of terrain, the vertical profiles of wind and exchange coefficients and the effects of non-stability are given in the last section.

The Diffusion of Radioactive Gases in the Mesoscale

(20 km - 400 km)

1. Introduction

The diffusion of radioactive gases in the atmosphere is a meteorological problem; one has to calculate the concentration of the gas (or gases) at a certain time after release and in a certain place, for a given source strength.

Once the field of the concentration (or the field of activity per unit of volume) in the lee of the source has been determined, all the other quantities required, such as the cloud dosage, rate of deposition, etc. can be calculated.

Quantitative methods of estimating diffusion have so far only been applied to the immediate area of the source, i.e. up to source distances of 10 km or at the most, 20 km. However, it is possible to obtain at least qualitative information for very large distances from the source, where $x > 300$ or 400 km, by calculating the trajectories of the "centres of gravity" of the dispersing gas clouds. The intermediate range between 20 km and 400 km referred to as the mesoscale (a term taken from meteorology), has so far been somewhat neglected in attempts to estimate diffusion. This paper attempts to fill this gap.

2. Delimitation of the mesoscale from the meteorological point of view

2.1. Atmospheric turbulence and diffusion

Dispersion in the atmosphere of substances not normally present in the air occurs by turbulent diffusion. Air flows are always turbulent and this means that diffusion is also always turbulent, although its intensity may vary. The relevant coefficient of turbulent diffusion k_s $\left[cm^2\ sec^{-1}\right]$ is 5 - 6 orders of magnitude larger than the coefficient of molecular diffusion, which could be used to describe the process of diffusion if air flows were laminar.

The dispersion of substances not normally present in air is thus determined by the nature of atmospheric turbulence, and it is, therefore, reasonable to define certain diffusion ranges in relation to turbulence.

Turbulence is classified by reference to a characteristic length (or "scale"), e.g. the vortex diameter, as :

microscale turbulence (also known as microturbulence),
mesoscale turbulence, or
macroscale turbulence (also known as large scale turbulence or macroturbulence).

Each of these types of turbulence is caused by different meteorological effects; and in each type the structure of turbulence is different, as is its ability to diffuse substances not normally present in air. This applies especially to the range in which diffusion takes place, and the three types of turbulence can be related to three ranges of diffusion, i.e. the

short range of diffusion,
medium range of diffusion, or
long range of diffusion,

with the same characteristic lengths ("scales") as those applied to the particular type of turbulence.

Before giving more concrete information on how the individual scales for turbulence are determined, it should be pointed out that this is somewhat problematic in that the characteristic lengths in vertical and horizontal directions are different. But as we are using turbulence conditions to refer to horizontal ranges of diffusion, the characteristic length in the horizontal plane will have to be considered.

A distinction is made between the following scales :

microscale	$<$ 20 km
mesoscale	20 – 400 km , and
macroscale	$>$ 400 km.

The two distances concerned, 20 km and 400 km, correspond to the following frequencies f /Hz/, or periods τ, depending on the (time-averaged) wind speed :

		5 m/s	15 m/s
20 km	f =	2.5 10^{-4} Hz	7.5 10^{-4} Hz
	τ =	1 hr 6 min 40 sec	22 min 13 sec
400 km	f =	1.25 10^{-5} Hz	3.75 10^{-5} Hz
	τ =	22 hr 11 min 40 sec	7 hr 24 min 26 sec

Table 1

Examination of the spectral density distribution of turbulent kinetic energy (turbulence spectrum) shows three different ranges of frequency: high energy levels are found in the microturbulence and macroturbulence ranges, while very little energy is found in the range of mesoscale turbulence, and we speak of a "gap" in the spectrum. The ranges can be easily subdivided on the basis of the frequencies given above (these vary not only with the averaged wind speed, but also with the height above the ground, and for this reason a non-dimensional frequency f' is normally used ($f' = fz/\bar{u}$)).

As waste gases disperse further and further away from a source, larger and larger eddies will be involved in the dispersion process, i.e. the frequencies of the turbulence spectrum affecting dispersion will be correspondingly lower. In this way all the eddies take part, in sequence, in the process of dispersion, beginning with the very high frequencies (in the dissipation range, i.e. molecular diffusion : initially, only microturbulence occurs, producing diffusion over the short range (<20 km), followed by mesoscale turbulence, which we know occurs with only relatively low energy levels (producing only very slight changes in the direction of transport of the suspended matter). Lastly, high-energy macroturbulence occurs and moves the diffusing matter to and fro, depending on the synoptic weather conditions, thereby causing diffusion over a wide area.

The delimitation of the scale at 20 km and 400 km is of course somewhat arbitrary; it would also have been possible to take the distances of 10 km or 25 km for the lower boundary and 200 km or 300 km for the higher. In the next part of the study, where the structure of atmospheric turbulence and the various alternatives for quantitative or qualitative attempts of estimates of diffusion are discussed, it will become clear that limiting the mesoscale at 20 km and 400 km is meaningful and justified.

2.2. Microturbulence

Microturbulence is generally caused by the (vertical) shear of horizontal wind in the atmospheric boundary layer. Production of turbulent kinetic energy is largest near ground level and decreases roughly exponentially with height. At the top of the boundary layer, i.e. at an approximate height of 300 - 500 m for a stable stratification and at 1500 - 2000 m for an unstable stratification, production of turbulence energy drops to zero. Microturbulence may also be caused by irregularities and roughness (vegetation, built-up areas, etc.) of the earth's surface.

Microturbulence is three-dimensional; the turbulent fluxes produced by microturbulence can be determined by the flux-gradient relationship in the same way as the molecular fluxes; this can be used in calculating diffusion.

As the characteristic length for microturbulence we use the asymptotic mixing length 100, obtained at the top of the boundary layer; it is 20 m for neutral stratification, but only a few metres for a stable stratification and ℓ_∞ m or more in unstable stratified boundary layers. Another characteristic length would be the diameter of the largest eddy occurring in microturbulence; vertically that corresponds to the thickness of the boundary layer, i.e. approx. 1 km for neutral stratification. In the horizontal direction one has to take for this a few kilometers depending on the degree of homogeneity of the underlying surface.

The reason why the distance of 20 km was chosen as the limit between the microscale and the mesoscale is that here the so-called convection-scale is included into the microscale. In meteorology there is a fourth range between the microscale and the mesoscale, the convective or convection-scale, due to thermal convection. In this scale we find all the (already more organized) processes caused by the boyant upward motion of air masses heated at the ground, eventually resulting in the formation of cumuliform clouds. In large cumulonimbus clouds these motions can extend up as far as the tropopause and the maximum vertical length should therefore be taken as roughly 10 km; the characteristic lengths involved in the horizontal direction are somewhat larger, and 20 km would appear to be a suitable figure.

2.3. Macroturbulence

Macroturbulence is generated in a completely different way from microturbulence. All eddies imbedded in a zonal flow are classed as macroturbulence; these include high and low pressure systems shown on weather maps and other disturbances studied in synoptic meteorology. (For this reason the macroscale is often referred to as the synoptic scale). These large-scale eddies are caused by baroclinic instability, and they are eddies with a vertical axis; unlike the eddies in microturbulence they are only two-dimensional. This means that the flux-gradient relationship can no longer be applied; doing this would involve negative (!) turbulent diffusion coefficients for the macroturbulent momentum flux, i.e. rather than diffusion or dispersion, this process involves "confusion" or increasing concentration of momentum.

2.4. Mesoscale turbulence

The relatively energy-free intermediate range between microturbulence and macroturbulence is taken up by mesoscale turbulence. This type of turbulence is primarily responsible for the diffusion in the mesoscale (20 - 400 km) which is being studied here.

It is originated by a number of different factors, some of which are given below. They include, firstly, deviations from main (or time-averaged) flows due to local thermal wind systems. Among these are land and sea breezes as well as mountain and valley winds.

As well as this, all small-scale synoptic phenomena take the form of mesoscale turbulence, such as air flow in weather fronts, lines of convergence and divergence, and larger convective formations. These include large cloud clusters with mesoscale dimensions, and which can extend as far as the macroscale in the tropics. As well as this, open convection cells occurring over sea areas, with sizes up to several hundred kilometers, belong in the mesoscale. But above all, there are the disturbances in the air-flow caused by changes in the topography of the earth's surface, which also occur, to a large extent, in the mesoscale.

It will, therefore, be seen that of all three types of turbulence (microturbulence, macroturbulence and mesoscale turbulence) the last is the most difficult to assess. In different instances it may have very different causes, a different structure and therefore different effects on the diffusion of matter. This is the main difficulty to be overcome in quantitative estimate of diffusion in the mesoscale.

3. Present state of methods to estimate diffusion in the microscale and macroscale

3.1. General

Unlike turbulence in the mesoscale, the structure of turbulence in the microscale and macroscale is known to some extent and understood at least as well for diffusion to be estimated. The results of such estimates are, however, quantitative only for the microscale, and can as yet be no more than qualitative for the macroscale. But no diffusion estimates of any kind are available for the mesoscale. Such estimates can only be obtained by extrapolation of the quantitative results available for the microscale beyond the 20 km limit or by extrapolation of qualitative results for the macroscale to below 400 km. Thus it is essential to begin by giving a short summary of the methods of estimating diffusion in the microscale (short range) and the macroscale (long range).

3.2. Quantitative methods to estimate diffusion in the microscale

Microturbulence is three-dimensional and can thus be described by analogy with molecular collision processes, treating dispersion caused by microturbulence as a molecular diffusion, however with the difference that the diffusion coefficient is larger (by 5 - 6 orders of magnitude). The concentration distribution of matter emitted continuously or instantaneously from a source can then be expressed in terms of normal distributions which can, in each coordinate direction, have completely different root mean squares σ_x, σ_y and σ_z.

However, some assumptions have to be made which considerably restrict the applicability of the results obtained by this method, viz.

(a) The meteorological parameters $\bar{u}$, σ_x, σ_y, σ_z are constant with time
(b) Horizontal homogeneity of these meteorological parameters
(c) Completely flat, even terrain
(d) The meteorological parameters must be constant with height; vertical mean values of $\bar{u}$ are used for this. The root mean squares σ_y and σ_z used in these calculations were determined empirically in prior diffusion experiments assuming that they were constant with height
(e) With continuous emission from source - which is the case most often studied - turbulent diffusion in the direction (x) of the mean wind is neglected as being very small compared with advection by the mean wind itself, i.e. $\sigma_x = 0$.

With these limitations, the concentration field of emission from an instantaneous or continuous source (at a given point) can be calculated, given:

(f) The source strength [g] or [Ci] for an instantaneous source or in [g sec^{-1}] or in [Ci sec^{-1}] respectively for a continuous source;

(g) The "effective" height of the source above ground level. Where an over-rise above the chimney is formed by hot effluent gases, further data are required to calculate the height of this over-rise,

(h) Deposition velocity on the earth's surface, which depends on the type of aerosol and the nature of the coverage of the ground;

(i) The radioactive decay rate;

(j) The following meteorological parameters: $\hat{\bar{u}}$, that is, the velocity of the horizontal wind averaged over the vertical coordinate ($\wedge$) and time ($-$), and σ_y (x) and σ_z (x), the height-constant root mean squares of the two normal distributions.

Also the direction of the vertically-averaged and time-averaged wind must of course be given; this is used to determine the direction of the x-axis.

Instead of the turbulent diffusion coefficients k_y and k_z, the appropriate values of σ_y and σ_z are used; these are the roots of the mean displacement squares of particles in $\pm$ y or $\pm$ z directions; the relation between k and σ is

$$2\, k_y \frac{x}{\hat{\bar{u}}} = \sigma_y^{\,2} \qquad\qquad 2\, k_z \frac{x}{\hat{\bar{u}}} = \sigma_z^{\,2} \qquad (1)$$

where x is the source distance ($x = \hat{\bar{u}}\, t$), increasing with travel time t. Using σ_y (x) and σ_z (x) has one great advantage: whereas the relevant σ belonging to Fick's diffusion with k as a constant is

$$\sigma_y \propto x^{0.5} \quad , \quad \sigma_z \propto x^{0.5} \qquad (2)$$

we know that at the beginning of diffusion, i.e. close to the source

$$\sigma_y \propto x^{1.0} \quad , \quad \sigma_z \propto x^{1.0} \qquad (3)$$

both (2) and (3) are recognisable by Taylor's theorem as limiting cases. a relationship for the dependence of σ_y and σ_z on source distance x may be obtained with the aid of the statistical theory of turbulence; in that, however, the Lagrangian autocorrelation function appears; this quantity cannot be measured in the atmosphere, that is why more or less realistic, and thus questionable, assumptions for it have to be made. For practical calculation of diffusion, therefore, it is advantageous to use dependencies σ_y (x) and σ_z (x), obtained empirically, in the binormal formula for concentration distribution $\bar{s}$ (x,y,z)

$$\bar{s}(x,y,z) = \frac{Q}{2\pi\bar{u}\;\sigma_y(x)\;\sigma_z(x)} \exp\left\{-\frac{y^2}{2\;\sigma_y(x)^2}\right\} \times \left[\exp\left\{-\frac{(z-h)^2}{2\;\sigma_z(x)^2}\right\} + \exp\left\{-\frac{(z+h)^2}{2\;\sigma_z(x)^2}\right\}\right] \qquad (4)$$

The concentration distribution (4) has already been somewhat simplified to allow for the fact that the deposition speed is not included; at ground level the total reflection condition, which is normally only applicable to gases, is used. h is the effective source height, Q [g sec^{-1}] is the strength of continuous emission from a point source; the x direction is that of the time-averaged horizontal wind $\vec{v} = \vec{i}\;\bar{u}$). Examples of this type of empirical dependence are given in PASQUILL's diagrams (1961); these show σ_y (x) and σ_z (x) for six different "diffusion categories", which differ primarily in thermal stratification and will be determined using conventional meteorological measurements (e.g. synoptic observations). PASQUILL's diagrams as given in "Meteorology and Atomic Energy" (Slade 1968) show values for σ_y (x) and σ_z (x) in the 100 m $<$ x $<$ 100 km range, i.e. extending into the mesoscale. But at the same time it should be pointed out that they contain only very vague extrapolations in the range x $>$ 20 km.

Diffusion of gases in the short range can be calculated using formula (4). The concentrations obtained differ from those actually measured by an average factor of 2, and are thus not very usable. These deviations must be due to the fact that in most cases the above-mentioned conditions (a), (b) and (c) are not fulfilled, that the vertical average $\hat{\bar{u}}$, as in (d) only gives a very imprecise equivalent height-constant value, and that the assumption in (e) is not satisfied at low wind speeds, (i.e. especially near ground level). Another important point is that the value for source strength is often very imprecise or, for continuous sources - and most of them are continuous - tends to fluctuate. In most cases the effective source height can only be calculated approximately, deposition velocity is almost unknown (and is thus assumed to be zero in calculations, i.e. total reflection from the earth's surface is assumed) and, lastly, the diffusion categories, and therefore the diffusion parameters σ_y (x) and σ_z (x) can only be very roughly defined. So it is not surprising that the results obtained by calculation and by actual measurement do not agree; but these imperfect results on short range diffusion are still the best mathematical statements available as yet.

3.3. Qualitative methods to estimate the diffusion in the macroscale

In the macroscale, i.e. in long range diffusion, it is only possible to calculate where matter released at a certain point will drift to; it is not possible to calculate the concentration of this matter when it arrives at a given point. Therefore we can only work out the trajectory of the centre of gravity of a cloud of gas, but this can be calculated for very large distances. For example, it is possible not only to trace the trajectory of a cloud of matter released by a bomb explosion up to once or twice around the earth, but to predict the trajectory, of course with an accuracy decreasing with distance.

Nowadays such calculations of trajectories are relatively straightforward, as the methods of "numerical weather prediction" provide us with information on all the required atmospheric fields in ten-minute time steps. By these methods fairly reliable large-scale predictions of atmospheric fields (horizontal wind w_h, vertical velocity w, atmospheric pressure p, temperature T and if necessary specific humidity q), for the next 48 hours, can be obtained. The system is based on numerical integration of a system of equations describing the conservation of mass, momentum and energy. These (non-linear partial) differential equations are replaced by finite difference equations and solution functions are then obtained for a large number (approx. 10000 - 15000) of grid points of a three-dimensional mesh. Because of the large amount of computer time this requires, choice of the mesh-width must be based on a compromise between one as small as possible, with high resolution, and one which is fairly large, i.e. with a minimum of grid points, and therefore requiring not so much computer time. In the prediction models currently used, the horizontal mesh-width is 350 - 400 km, with a vertical mesh-width of 200 mb or more. This indicates the degree of resolution.

It will be seen that the lower limit of resolution corresponds to the boundary between the mesoscale and the macroscale, and that numerical models for weather prediction can therefore only predict large-scale fields of atmospheric variables, i.e. synoptic fields. This means that the vertical motions obtained by these models are also only those for the large scale; vertical mixing within the troposphere by strong thermal convection (cumulonimbus clouds) does occur in the mesoscale and therefore cannot be covered by these models. The topography of the terrain is smoothed depending on the mesh-width chosen, and only very high mountain ranges can be considered; they are treated as if they were completely smooth, without any valley rifts. The winds flowing over mountain ranges of this type cause additional (large-scale) vertical motions. But these models cannot be used to calculate mesoscale effects which are most important for the diffusion of released matter - such as the channelling of air flows through valleys, the formation of luff and lee eddies in front of or behind large mountain ridges, and other phenomena.

With these models, only large-scale trajectories can be computed, giving the three-dimensional displacement of the centre of gravity of a diffusing cloud of released matter caused by the air flow in the macro-scale.

The most highly developed atmospheric models for numerical weather prediction are not, like most models, "dry", but incorporate predictions of the field of specific humidity; this includes predictions of rainfall. Because of this there is a possibility that the wash-out effect can be determined along a trajectory (if only very approximately); that would represent a step towards quantitative, and away from qualitative, estimates of diffusion.

4. Current state of the ability to estimate the diffusion in the meso-scale

4.1. Climatological information on the probability of impact

Often the frequency of wind in certain directions (wind rose), obtained by climatological observations is used to calculate the probability of which radioactive gases will move in a certain direction after an assumed reactor accident. This would only apply if reactor accidents would occur very frequently at the same place. But only when a particular direction is found in the vast majority of all cases e.g. for reasons connected with orographic features (such as channelling of air flows through a valley), can approximative predictions of the probability be made.

This is of course with the assumption that the wind direction during the diffusion-process over the diffusion range is the same as at the source where the wind rose was recorded. The deviations of flow directions along the trajectory, e.g. by mountain ridges or channelling effects in valleys, which alter the direction, and other factors, are not considered. Thermal stratification in the atmospheric boundary layer is not taken into account either; but it is nevertheless possible, on the one hand, for thermal convection to produce total vertical mixing throughout the troposphere, where the matter transported upwards is displaced in a completely different horizontal direction from that at ground level, and, on the other hand, it is possible for the whole diffusion process to be restricted to the lowest layer because of the existence of an inversion "lid".

All in all, the use of wind roses cannot really be considered as a reliable method of obtaining probabilities.

4.2. Bibliographical survey on quantitative estimates of the diffusion in the mesoscale

This section lists the available papers which make some attempt at quantitative estimates of the diffusion of matter in the mesoscale ($20\ \text{km} < x < 400\ \text{km}$). The list does not include papers which only give qualitative data, e.g. collections of data on concentration and deposition of radioactive matter after the Windscale accident, in publications such as those by STEWART and CROOKS (1958) and by CHAMBERLAIN and DUNSTER (1958).

The papers are listed in chronological order. A brief note on each paper is included, showing the most important quantitative result it gives, usually expressed in terms of the distance dependence of the diffusion parameters $\sigma_y(x)$ and $\sigma_z(x)$.

All in all it must be said that only very few data are available which are really reliable and founded on sufficient observations. The major part of the papers mentioned use the diffusion parameters $\sigma_y(x)$ and $\sigma_z(x)$ based on PASQUILL's diagrams and the "experimental" basis of most of the calculation methods given below is thus the same.

1932 SUTTON, O.G.

A Theory of Eddy Diffusion in the Atmosphere
Proc. Roy. Soc. A, 135, 143

Reanalysis of the data obtained by RICHARDSON and PROCTOR (1926) with the result

$\sigma \propto x^{0.875}$ $50\ km < x < 500\ km$

No distinction with respect to the thermal stratification, all cases taken together.

1960 MARTIN, J.J.

Etude des consequences de l'accident de Windscale (October 1957) et de la validité du modèle mathématique de diffusion atmosphérique de Sutton
C.E.N. Saclay, Rapport CEA-1538, 31 pp.

Application of Sutton's diffusion formulae to measurements after Windscale accident.

$\bar{s}\ (x, 0, 0) \propto x^{-(2-n)}$ or with $n = 0.2$

$\sigma_y \cdot \sigma_z \propto x^{1.8}$

1962 PASQUILL, F.

Atmospheric Diffusion
D. Van Norstrand Comp. Ltd, London
297 pp.

p. 169 : Evaluation of Porton data and those of BRAHAM, SEELY and CROZIER (1952)

$\sigma_y \propto x^{0.8}$ $3\ km < x < 80\ km$

p. 209 : Empirical diagrams

$\sigma_z \propto x^{0.81}$ Category D_1 neutral stratification

$\sigma_z \propto x^{0.23}$ Category F stable stratification

both for $10\ km < x < 100\ km$

$\sigma_y \propto x^{0.84}$ Category D_1 and F

(last determined by diagrams in SLADE 1968)

1962 PASQUILL, F.

Some Observed Properties of Medium Diffusion in the Atmosphere
Quart. J. Roy. Meteor. Soc 88 (375), 70-79

Evaluation of tracer experiments in the range $15\ km < x < 150\ km$ with regard to longitudinal and lateral diffusion; but no direct evaluation of $\sigma_y\ (x)$ and $\sigma_z\ (x)$

1963 PACK, D.H.
ANGELL, J.K.

Air Trajectories and Turbulence Statistics from Weather Radar Using Tetroons and Radar Transponders
Monthly Weather Rev. 91 (10/11), 583-604

A summarizing diagram available in SLADE (1968), Fig. 4.44 p. 180. From this one gets

$\sigma_y \propto x^{0.85}$ $\qquad 10 \text{ km} < x < 50 \text{ km}$

6 flight series in the Los Angeles basin are analysed, with of course different thermal stratifications.

1964 BRYANT, P.M.

Methods of Estimation of Windborne Material and Data to Assist in Their Application
U.K. - A.E.A., Report AHSB (RP) R. 42, H.M.S.O.

Data on diffusion in the range 10 km - 100 km, mainly based on PASQUILL's diagrams.

1965 HEFFTNER, J.L.

The Variation of Horizontal Diffusion Parameters with Time for Travel Periods of One Hour or Longer
J. Appl. Meteor 4 (1), 153-156

Collection of measurement data and analyses by 19 authors (range 25 km < x < 10,000 km)

Conclusion : mean $k_y = 4.10^8 \text{ cm}^2 \text{ sec}^{-1}$; i.e. acc. eq. (1) : $\sigma_y \propto x^{0.5}$

For the mesoscale these data would suggest (although the author does not state this)

$\sigma_y \propto x^{0.89}$

1965 SMITH, M.E.
SINGER, I.A.

An Improved Method of Estimating Concentrations and Related Phenomena from a Point Source Emission.
USAEC Rep. BNL-9700, Brookhaven National Laboratory

Results are summarized in diagram form in SLADE (1968) p. 128 :
BNL-trace-type D for stable stratification is given as

$\sigma_y \propto x^{0.72}$ $\qquad 10 \text{ km} < x < 70 \text{ km}$

for other stratifications trace types are only given for the short range ($x < 10$ km).

1968 ISLITZER, N.F.
SLADE, D.H,

Diffusion and Transport Experiments
Chap. 4 in "Meteorology and Atomic Energy 1968" SLADE, D.H. (Editor)
USAEC Div. Techn. Inform. Oak Ridge, Tenn., USA, 445 pp.

p.132 : Statement that σ_y-summary diagram (Fig. 4.21 on p. 148) shows

$$\sigma_y \propto x^{0.85}$$

The analyses shown in this diagram extend to approx. x = 30 km into the mesoscale

p. 182 : The summary diagram (Fig. 4.45) indicates, that according to an evaluation of tetroon flights

$\sigma_y \propto x^{0.87}$ $\quad$ 5 km < x < 35 km
(taking flights together over a period of 24 hours) and

$\sigma_y \propto x^{0.84}$ $\quad$ 5 km < x < 35 km
(taking flights together over a period of 6 hours only)

1969 HILST, G.R.

Simulation Model for the Air Pollution over Connecticut
Travelers Research Corporation,
Rep. Nr. TRC 7242 a,b (2 Vols.)

1969 IZRAEL, U.A.
PETROV, V.N.
PRESSMAN, A.A.
ROVINSKY, F.A.
STUKIN, E.D.
TER-SAAKOV, A.A.

Radioactive Contamination of the Environment by Underground Nuclear Explosions, and Methods for Forecasting It
USAEC - Translation 7122

After the Soviet test explosion "1003", radiation r was studied in relation to crater distance along the axis of the fall-out area. Data cover the mesoscale

$r \propto x^{-2.14}$ $\quad$ for "1003"

which may be compared with the result of two test explosions in the West :

$r \propto x^{-2.0}$ $\quad$ for "Danny Boy"

$r \propto x^{-2.38}$ $\quad$ for "Sedan"

From this it is possible to calculate diffusion parameters (with certain additional assumptions) e.g. the product of $\sigma_y \cdot \sigma_z$.

1969 TURNER, D.B.

Workbook of Atmospheric Dispersion Estimates US-Department of Health, Education and Welfare, NAPCA Cincinnati Ohio, 84 pp.

Diagrams for σ_y (Fig. 3-2) and σ_z (Fig. 3-3) for diffusion categories A - F, e.g.:

C	$\sigma_z \propto x^{0.96}$	$\sigma_y \propto x^{0.88}$
D	$\sigma_z \propto x^{0.68}$	$\sigma_y \propto x^{0.88}$
F	$\sigma_z \propto x^{0.31}$	$\sigma_y \propto x^{0.88}$

all for 10 km $< x <$ 100 km

1970 BEATTIE, J.R.
BRYANT, P.M.

Assessment of Environmental Hazards from Reactor Fission Product Releases UK-AEA Report AHSB (S) R 135 HMSO, 24 pp. + 34 Diagrams

Large number of diagrams, based on PASQUILL's diffusion parameters for categories C and D on the one hand, and F on the other; additional assumptions for deposition velocity and rain-out effect.
Examples: (Fig. 4) Cloud dosage at ground for continuous release, category F

$CD \propto x^{-1.32}$ $\quad v_d = 0$

$CD \propto x^{-1.98}$ $\quad v_d = 3.0$ mm sec^{-1}

for 10 km $< x <$ 100 km

1970 BULTYNCK, H.
MALET, L.
SHARMA, L.N.
VAN DER PARREN, J.

Atmospheric Dilution Factors and Calculation of Doses in the Environment of S.C.K./C.E.N. MOL for Short and Long Duration Stack Discharges S.C.K./C.E.N. MOL, Report BLG 446, Part I 83 pp., Part II 58 Fig.

Calculations only up to 10 km, except for the fumigation type; and for some correction factors extrapolation up to 50 km

$\bar{s}\,(x,0,0) \propto x^{-0.80}$

1970 LE QUINIO, R.

Evaluation de la diffusion d'effluents gazeux en atmosphère libre à partir d'une source ponctuelle continue; abaques et commentaires C.E.N. Saclay, Rapport CEA-R-3945 Février 1970, 20 pp.

The diagram for $\bar{s}$ (x,0,0) extends from the short range to 50 km for unfavourable conditions, and to 25 km for normal diffusion conditions. It uses for the first :
$\bar{s}(x,0,0) \propto x^{-1.38}$ i.e. $\sigma_y \cdot \sigma_z \propto x^{1.38}$
but for the latter :
$\bar{s}(x,0,0) \propto x^{-1.20}$ i.e. $\sigma_y \cdot \sigma_z \propto x^{1.20}$
If we take the same variation $\sigma_y \propto x^{0.88}$ for unfavourable and normal dispersion conditions, in the first we would have
$\sigma_z \propto x^{0.5}$ and $\sigma_z \propto x^{0.32}$ for the latter. There must be an error here, as diffusion cannot be less intensive in normal conditions than in unfavourable conditions.

1971 BERLYAND, O.S.
PETROV, V.N.

The Effect of Atmospheric Variables in the Diffusion and Settling of Radioactivity from Clouds Travelling over Long Distances
Bull. (Izv.) Acad. Sci. USSR, Atm. Ocean. Phys. 7 (11), 1209-1214 (Engl. transl.)

Theoretical study: Use of one-dimensional (!) time-dependent Fickian diffusion equation (implying that $\sigma_z \propto t^{0.5}$); transformed to source distance. Study of fall-out assuming deposition velocity and rain-out effect

1971 O.E.C.D.
Environment Directorate
Study Group

Models for Prediction of Air Pollution
Paris, 15 April 1971, 80 pp.

Lists all requirements for models of various kinds:
p. 53 : Diffusion for $x > 20$ km,
p. 54 : Diffusion for $x > 100$ km
However no data suitable for practical application

1972 CAGNETTI, P.
PAGLIARI, M.

Long Distance Transport and Diffusion of Gaseous Effluents: Considerations and Calculations for a Health Physics Analysis

Definition of the mesoscale as 20 km $< x <$ 500 km (p.9); asymptotic values for σ's based on HEFFTNER's (1965) data
$\sigma_y \propto x^{1.5}$ (p.9), but $\sigma_y \propto x^{1.25}$ (p.13)
at high wind speeds

1972 BERLYAND, O.S.
PETROV, V.N.
SEVEROV, D.A.

Propagation of a Contaminant over Long Distances in the Case of a Vertical Diffusion Coefficient That Varies with Time or Height
Bull. (Izv.) Acad. Sci. USSR, Atm. Ocean. Phys. 8 (9), 994-997
(Engl. Transl. 575-577)

Extension of BERLYAND and PETROV's (1971) work; k_z is determined by measurements, but sinusoidal diurnal variation is assumed.

1972 DOURY, A.

Une méthode de calcul pratique et général pour la prévision numérique des pollutions véhiculées par l'atmosphère
C.E.N. Saclay, Rapport CEA-R-4280, Février 1972, 37 p.p.

A formula is given for non-Fickian diffusion, in which the following parameters are taken (without foundation)

$$\sigma_y \propto x^{1.13}, \quad \sigma_z \propto x^{0.50}, \quad x < 60 \text{ km}$$

$$\sigma_y \propto x^{1.00}, \quad \sigma_z \propto x^{0.50}, \quad x > 60 \text{ km}$$

1972 GROUPE COMMUNE
C.E.N. - VINCOTTE

Pollution résultant des effluents des cheminées - Méthode de calcul.
Report 03/041 Avril 1972

Applies bi-normal distribution, no statement on restriction in x-direction. Parameters chosen:

$$\sigma_y \propto x^{0.796}, \quad \sigma_z \propto x^{0.711}$$

for stability categories E 1 - E 6 and

$$\sigma_y \propto x^{0.698}, \quad \sigma_z \propto x^{0.669}$$

for inversion conditions (category E 7)

1973 CLARKE, R.H.

The WEERIE-Program for Assessing the Radiological Consequences of Airborne Effluents from Nuclear Installations
Health Physics 25, 267-280

Describes a computer programme used to calculate the concentration field and other variables derived from this; the programme is based on PASQUILL's diffusion parameters

1973 LYONS, W.A. OLSSON, L.E. Detailed Mesometeorological Studies of Air Pollution Dispersion in the Chicago Lake Breeze
Monthly Weather Rev. 101 (5), 387-403

Studies on the effect of horizontal and vertical non-homogeneity in the windfield on diffusion up to 40 km. Quantitative data on the flow field only, not on the concentration field.

4.3. Extension of diffusion estimates from microscale to mesoscale

Retaining bi-normal concentration distribution as the solution of the Fickian diffusion equation in the mesoscale as well - and at present no other means of quantitative estimate is available - one needs the two diffusion parameters σ_y (x) and σ_z (x) depending on source distance in the mesoscale. Beyond $x > 20$ km certainly the point has been reached where according to the theory, σ_y (x) and σ_z (x) vary only proportionally to $x^{0.5}$, as in (2), i.e. where turbulent diffusion acts physically in the same way as molecular diffusion. (Eddies of all possible sizes are involved to carry out the diffusion process, and the Lagrangian autocorrelation function for the velocity fluctuations has dropped to zero). If this were the case quantitative estimates of diffusion in the mesoscale would be extremely simple, but observation shows that σ_y and σ_z do in fact increase more rapidly than by $x^{0.5}$ and that because of this the concentration on the axis $\bar{s}$ (x,0,h) or on its projection at ground level $\bar{s}$ (x,0,0) decreases more rapidly with increasing source distance x than by $x^{-1.0}$. There are several reasons for this:

(a) the diffusion parameters σ_y (x) and σ_z (x) are determined by measuring the concentration in the lee of a source of known strength; the determination is based on bi-normal concentration distribution, which is obtained from Fickian diffusion equation for vanishing deposition velocity v_d. But if, as for the most concentration measurements, a deposition velocity exists other than zero, the concentration on the axis $\bar{s}$ (x,0,h) must decrease more rapidly than with vanishing deposition velocity; this is then reflected in a more rapid increase in $\sigma_y(x)$ and $\sigma_z(x)$ than by $x^{0.5}$.

(b) $\sigma_y(x)$ and $\sigma_z(x)$ are determined from measurements on the basis of the formulae mentioned in (a) which uses as condition at the lower boundary the total reflection of particles reaching the earth's surface. But if the earth's surface retains some of the matter being diffused (e.g. absorption by vegetation), this must be shown in turn by an increase in σ_y and σ_z with increasing source distance x, which is higher than by $x^{0.5}$.

(c) The greater the source distance x, the longer is the travel time of particles and the more noticeable the non-steadiness, which is incompatible with use of the aforementioned formula. This implies that the mean wind $\bar{v}$ will experience changes of direction and magnitude during the travel time of the particles, and will thus create increased lateral horizontal exchange; this implies that $\sigma_y(x) \propto x^p$ where $p > 0.5$.

(d) The greater the source distance x, the less satisfactorily the requirement of horizontal homogeneity is fulfilled. This is particularly true in the case of unstable stratification: instead of the small-scale irregular upward motion of warm air which determines σ_z in the microscale, the trajectories of diffusion particles can, in the mesoscale, be taken into organized systems of thermo-convection (e.g. entrainment into a cumulonimbus or cumulus congestus) and removed from the atmospheric boundary layer. This matter will then be found somewhere in the troposphere, it will not take any further part in the process of diffusion in the atmospheric boundary layer. Concentration thus decreases much faster with increasing source distance, i.e. $\sigma_z(x)$ must increase more rapidly than by $x^{0.5}$.

(e) In the experiments where $\sigma_y(x)$ and $\sigma_z(x)$ were determined by measurements of concentration, it was assumed that the diffusing matter did not undergo any physical or chemical change and that the total mass released at the source was conserved. The greater the source distance, the more likely it is that there will be some loss, caused for example by chemical reactions, or by decay in the case of radioactive matter (which of course can be treated quantitatively). Reactions of this kind also lead to a situation where $\sigma_y(x)$ and $\sigma_z(x)$ increase more rapidly than by $x^{0.5}$.

(f) The greater the source distance, the less the assumption of a flat earth surface is likely to be satisfied. Whenever the diffusing matter passes over rising ground with the airflow, the deposition on the luff side of this is increased. This has the same effect on the variables $\sigma_y(x)$ and $\sigma_z(x)$ as the points mentioned above in (a) - (e).

This is why empirical measurement of $\sigma_y(x)$ and $\sigma_z(x)$ will always show a large scatter, caused by the factors mentioned in (a) - (f), which may be completely different in different cases. As an average value (based on the different data given by all authors) we may, for the meso-scale, take $\sigma_y(x) \propto x^{0.8}$, whereas the data for $\sigma_z(x)$ must be discriminated according to thermal stratification.

One other point should be made about $\sigma_z(x)$: PASQUILL's diagrams (1962) give $\sigma_z(x) \propto x^{0.81}$ for diffusion category D_1, but $\sigma_z(x) \propto x^{0.23}$ for diffusion category F. The latter is theoretically not possible as with bi-normal distribution the lower limiting case would correspond to $\sigma_z(x) \propto x^{0.50}$, i.e. Fick's diffusion. But when experiments give an exponent lower than 0.50, this is usually because the bi-normal concentration distribution cannot be applied to the cases grouped in diffusion category F. These are mostly cases with a temperature inversion, which acts as a barrier against any upward diffusion. If diffusion were blocked completely, σ_z would be a constant independent of x, and the exponent would thus be 0; an exponent obtained from experiments, such as PASQUILL's value of 0.23 for instance, is more or less accidental and has little significance, depending as it does on the cases which happen to be considered in the evaluation and, unfortunately, these are always very few.

4.4. Estimating the dilution in the mesoscale in particularly unfavourable conditions

As described above, the most unfavourable case is one where further diffusion upwards is blocked by a temperature inversion at a certain height D. Beyond a certain distance x_D, the concentration then will be found to be constant with height in the layer $0 < z < D$. In such cases a dilution of concentration with increasing source distance can only occur by horizontal (lateral) diffusion, and will therefore be determined by the behaviour of $\sigma_y(x)$ only. Assuming the most unfavourable case for this, i.e. one with the smallest possible p in the proportionality $\sigma_y(x) \propto x^p$, it is possible to state the maximum dilution of concentration in the mesoscale. This consideration forms the basis of the statement given below.

The bi-normal concentration distribution is expressed as follows, when the (effective) source height h and the effect of total reflection at the earth's surface are disregarded:

$$\bar{s}(x,y,z) = \frac{Q}{2\pi\,\sigma_y(x)\cdot\sigma_z(x)} \exp\left\{-\left(\frac{y^2}{2\sigma_y(x)^2} + \frac{z^2}{2\sigma_z(x)^2}\right)\right\} \tag{5}$$

Now the part normally distributed over the vertical has to be equally distributed over the layer of thickness D

$$\int_{-\infty}^{+\infty} \exp\left\{-\frac{z^2}{2\sigma_z(x)^2}\right\} dz = \sqrt{2\pi}\;\sigma_z = D \tag{6}$$

so that

$$\begin{aligned} \bar{s}(x,y,z) &= \frac{Q}{\sqrt{2\pi}\;\sigma_y(x)\;D} \exp\left\{-\frac{y^2}{2\sigma_y(x)^2}\right\} \quad 0 < z < D \\ \bar{s}(x,y,z) &= 0 \qquad\qquad\qquad\qquad\qquad\qquad z > D \end{aligned} \tag{7}$$

Regardless of the height h (within layer D) at which the continuous source is situated, the matter released will first of all have to move over a distance x_D, until the concentration does not longer vary with height in the layer of thickness D. This distance x_D must be defined. According to equation (6)

$$\sigma_z = 0.4\ D \tag{8}$$

this means, that on the axis, i.e. at source height h, the concentration for normal distribution equals that one which is constant with height, when σ_z (x) has attained the value 0.4 D. The source distance x at which this occurs is then taken as the point (x_D) beyond which there is complete mixing in layer D, i.e. the concentration is constant with height.

$$x_D = x\ (\ \sigma_z = 0.4\ D\ )$$

PASQUILL's (1962) σ_z-values for diffusion category F are used to determine x_D. The following dependence is obtained.

D	$\sigma_z(x_D)$	x_D
100 m	40 m	1.0 km
200 m	80 m	2.3 km
300 m	120 m	3.7 km
400 m	160 m	5.3 km
500 m	200 m	7.0 km
600 m	240 m	8.8 km
700 m	280 m	10.3 km

Table 2
Dependence of distance x_D on the thickness of layer D

This dependence is illustrated in Fig. 1 on p. 27.

It may be seen that in a completely mixed layer with a lid caused by temperature inversion at 600 - 700 m height, the concentration is constant with height only beyond 10 km from the source.

Any data on the dilution of concentration in the mesoscale must be based therefore on the concentrations measured (e.g. at ground level) 10 km from a source; by doing this it is possible to allow for virtually

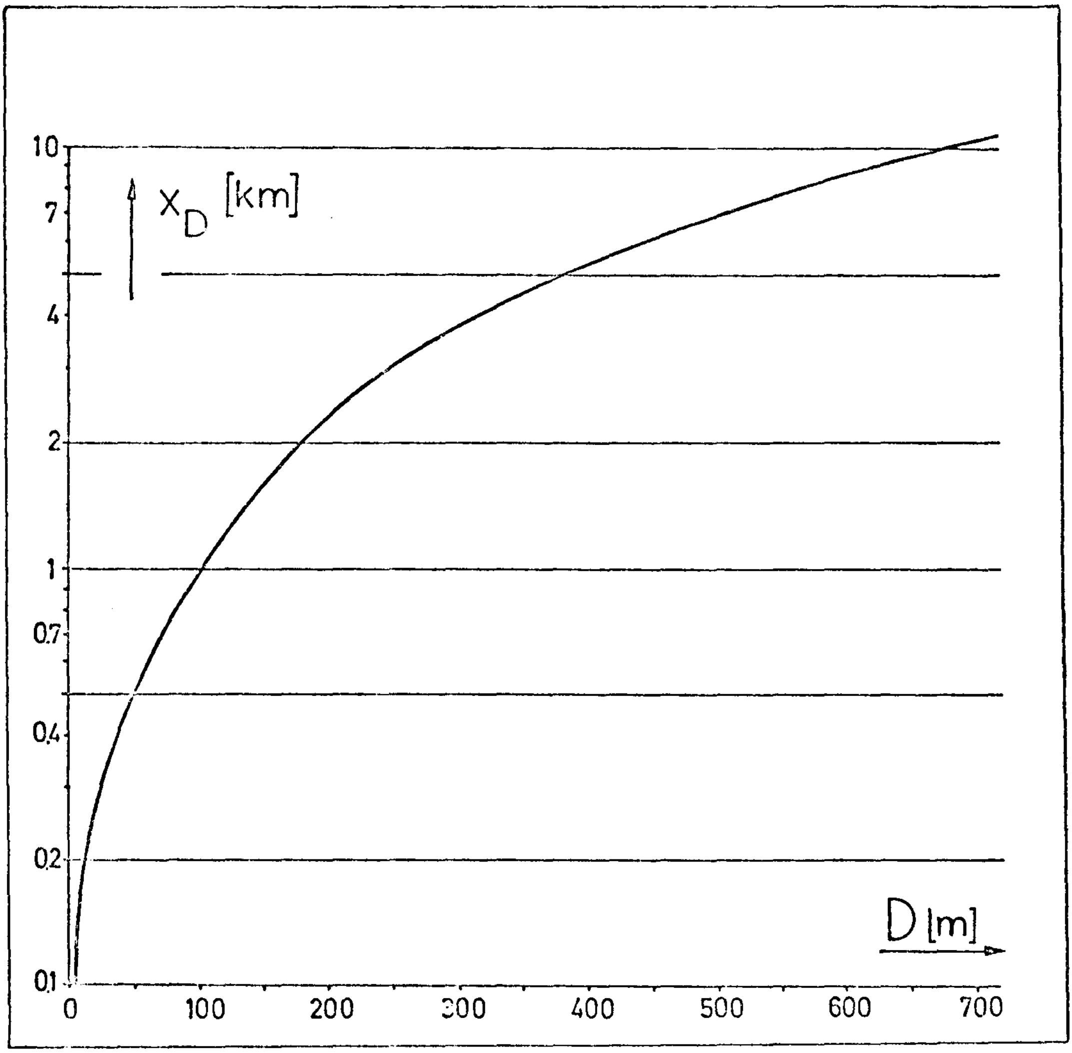

Fig. 1

Dependence of distance x_D on the thickness D of the mixed layer. Distance x_D is the distance beyond which the concentration (released at $x = \sigma$, $z = h$; $h < D$) does not vary with height.

any inversion below 700 m acting as a lid; using ground-level measurements of the concentration in the immediate vicinity of the source would very probably give values which are too low for mesoscale estimates, as the concentration still shows a distinct vertical profile with a maximum at height h of the effective source.

Following a nuclear reactor accident it would in any case be impossible to give an exact estimate of the quantity of noxious gases which have been released; instead measurements would have to be taken in the immediate vicinity, to know how seriously this area had been polluted, and then at a distance of 10 km, to obtain a reference value for the mesoscale.

If the following data are to be used to estimate dilution in the mesoscale during normal functioning of the reactor, the (continuous) emission Q $[g\ sec^{-1}]$ or $[Ci\ sec^{-1}]$ of noxious gases is known and the concentration $\bar{s}$ (10 km, 0 , 0 ; h) can be obtained from the usual diagrams e.g. PASQUILL (1962), and estimates on the dilution in the mesoscale can then be obtained by reference to this value.

In order to be able to estimate the dilution in the mesoscale, also an assumption must be made for the value of σ_y (x) in equation (7) for this range. This σ_y (x) should also cover the most unfavourable conditions. The value

$$\sigma_y\ (x) \propto x^{0.67} \qquad (10)$$

was selected. This choice is admittedly quite arbitrary, and cannot be fully substantiated. On the other hand it can be shown that this does in fact cover the most unfavourable conditions. Figure 2 (page 29) shows the σ_y-values for the mesoscale and large scale given by HEFFTNER (1965): the solid lines on the diagram (except for those marked with a W) are also shown by HEFFTNER, and connect points which were obtained in the same experiment. The curves P_A and P_F from PASQUILL's diagram for diffusion categories A and F are also shown, as broken lines, and the curve

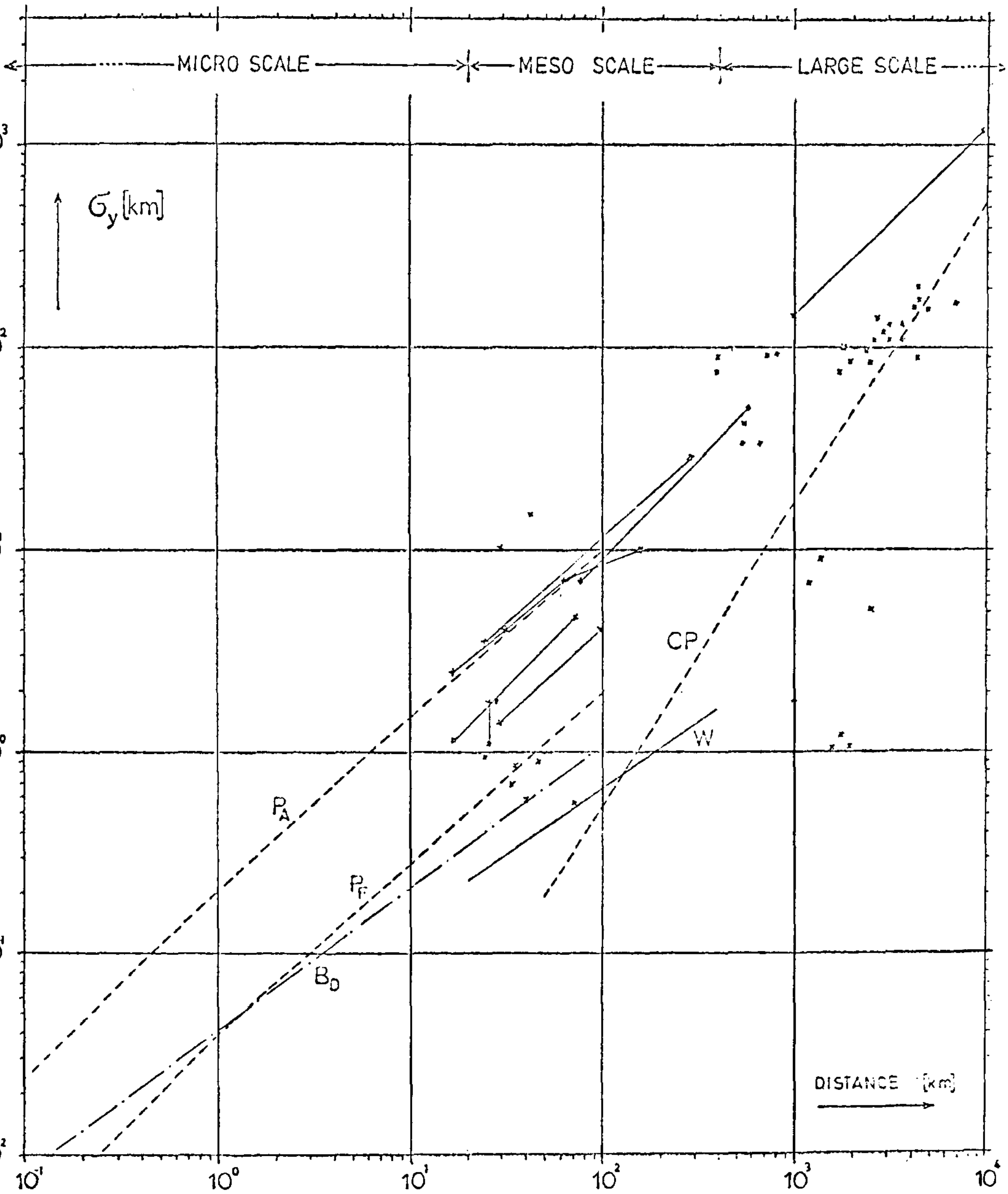

Fig. 2

Diagram of HEFFTNER's (1965) data for fixing the relation (10); this is the solid line marked W.

obtained for stable conditions (category D) from the Brookhaven measurements is shown as a dash-dotted line marked B_D. The broken line marked CP shows the dependence of σ_y (x) according to CAGNETTI and PAGLIARI (1972).

When the aim is to provide estimates on the dilution of matter in the mesoscale, the level of the curves in figure 2 is insignificant; the important factor is slope of the curves, i.e. the exponent p in the proportional expression $\sigma_y (x) \propto x^p$. This is for the individual curves given in table 3

curve	p
CP	1.21
P_F	0.85
B_D	0.71
W	0.67

Table 3

The curve CP is proposed for travel times of the particles; $(x/\hat{u})^p$ was substituted for t^p. Depending on the mean wind velocity assumed here, the curve only extends (from the macroscale !) partially into the mesoscale. The important slope of the curve is however unaffected by the wind speed assumed. The proposed curve CP would only apply under extremely favourable diffusion conditions,and can definitely not be applied for estimating dilution under unfavourable conditions. The reason why the exponents p for P_F and B_D are higher than for W, indicating that in these two cases concentration decreases more rapidly with increasing source distance, is probably that P_F and B_D is an average over all extremely stable situations, whereas W should represent a limiting case.

Where p = 0.67, a dilution of concentration $\bar{s}$ (x,0,0) with increasing x, as shown in Table 4, is obtained; concentration is normalized with those concentrations measured at 10 km distance (here care must be taken that these measurements are made in the precise direction of the mean wind $\bar{w}$, so as to obtain the concentration along the axis, y = 0.

x [km]	$\frac{\bar{s}(x,0,0)}{\bar{s}(10\text{ km},0,0)}$
10	1.0
20	0.59
25	0.51
50	0.32
75	0.25
100	0.20
125	0.18
150	0.16
175	0.14
200	0.13
250	0.11
300	0.10
350	0.09
400	0.08

Table 4
Minimum dilution of concentration $\bar{s}(x,0,0)$ with increasing source distance in the mesoscale
(concentrations related to measurements at 10 km distance)

Here the decrease of concentration between 10 km and 20 km was calculated with p = 0.75, as this range is still within the microscale; the value selected is between those of P_F and B_D.

For comparison, we give the dilutions at x = 100 km and x = 400 km which would be obtained if B_D and P_F were extrapolated into the mesoscale; dilution with CP is also calculated:

x	W	B_D	P_F	CP
100 km	0.20	0.19	0.14	0.03
400 km	0.08	0.07	0.04	0.004

Table 5
Dilution between 10 km and 100 km/400 km assuming various exponents p

4.5. Applying the calculation of trajectories as used in the macroscale to the mesoscale

As described in 4.3 and 4.4 the available (quantitative) methods of estimating diffusion in the microscale were applied to the mesoscale. In the same way it is possible to apply the methods used for estimates in the macroscale to the mesoscale, but in this case only qualitative estimates can be made. Section 3.3 showed how the trajectory of the centre of gravity of a dispersing gas cloud in the macroscale could be calculated, and explained that this is best done by applying one of the models used for numerical weather prediction. The resolution of these models is dependent on the horizontal mesh width of the grid used; this is approx. 350 - 400 km.

To calculate trajectories in the mesoscale, it is necessary to use a smaller mesh width. Because this calls for a considerable increase in computing capacity, it can of course only be done for a restricted area (approx. 2000 km X 2000 km) and cannot be global, or at least hemispherical, like the large scale models. In fact, methods are being developed, and have to some extent already been devised, to incorporate small areas with a reduced mesh width of, say, 150 km, 100 km, 80 km and even 50 km in the larger area (e.g. a hemisphere) of the numerical weather prediction model with horizontal grid width of 350 - 400 km. A "nest" is created in this process, referred to as nesting. In this process the boundary values of the nest are obtained by computations in the large area; using these boundary values, which do of course vary with time, it is possible to carry out computations for the smaller area which are more precise, because the resolution is better.

These calculations within smaller areas do of course call for a less "smoothed" topography of the earth surface; this will allow for highlands, but not for valleys. Even valleys as wide as the Upper Rhine rift valley are almost smoothed out by an overlapping mean with a characteristic length of 50 km (or 100 km or 150 km). Air currents would, however, be affected by the mesoscale topography of this smaller area.

There are other mesoscale effects which cannot be incorporated, even if the mesh width is reduced to 50 km. These include systematic thermal convection. For example, if the dispersing gas cloud is entrained into a huge cumulonimbus and transported upwards in this, possibly up to the tropopause, this cannot be simulated in any nesting model available at present. It should be pointed out, however, that increased efforts are now being made to obtain "parameterizations" of thermal convection, and there may well be a possibility of incorporating this effect in the model in the near future. This would mean that, at the very least, probability statement on upward dislocation resulting from thermal convection could be given for a mesoscale trajectory.

It is also possible to build a smaller nest within the first nest, with a horizontal mesh width as small as 20 km or even 10 km. This is also being studied, and is referred to as "telescoping". Although such methods are still only at the experimental stage, it is expected that they will also be available for use in the near future.

With this telescoping technique it would be possible to cover the whole of the mesoscale, taking the macroscale at one end as a starting point; using a mesh width of 20 km would even provide a link with the microscale at the other end.

5. Some suggestions for present possible improvements in methods of estimating the diffusion in the mesoscale

These suggestions only refer to improvements which are feasible, given the current state of our knowledge and the technical means available.

5.1. Influence of the form of the terrain

5.1.1. Topography of the earth's surface

A technique developed in France (Prof. FACY, Paris) using a terrain model in a flow channel, provides a means of determining ground level concentration in the lee of a continuous source point, as the fluid eats into different layers of colour on the terrain model, depending on the concentration of a specially prepared additive released at the source. The transitions from one colour layer to another, laid bare by the action of the fluid, correspond to the isolines of concentration; these can be calibrated accordingly, although the duration of the effect must be taken into account. If the terrain model were set up in such a way that it could be turned around a vertical axis in the flow channel from experiment to experiment, it would be possible to obtain terrain-influenced concentration distributions for different flow directions, i.e. for different wind directions (at the source). By evaluating the intensity of the effect ($\approx$concentration) for each point in the terrain and relating it to the frequency of the corresponding wind direction, it is possible to obtain a field representation showing the areas which are subject to particular risk because of topographical factors. A picture of this kind, showing the concentration as a function of the product wind direction x relative frequency, and of source distance, would be a great deal more valuable than conventional wind roses for predicting the probability of a certain area being affected; the part played by valleys and hilly terrain would be taken into account. Because of this it would in many cases give a substantially different picture from that given by the corresponding wind rose.

A similar project is currently being sponsored by the U.S. Environmental Protection Agency (EPA), using a terrain model representing the topography of an area in the State of Pennsylvania.

5.1.2. Ground coverage

According to assumption (c) in section 3.2. the earth's surface must be completely flat and even, i.e. must not be built up or covered by vegetation. This assumption, required for the quantitative estimation technique in the microscale, must of course also be fulfilled in the mesoscale, if the technique is to be extended to this range (see section 4.3) The effect of buildings and vegetation is to create "roughness" on the earth surface, expressed by a roughness-length z_o. This roughness-length has as yet not been taken into account at all in calculating diffusion,

A method described by WIPPERMANN and YORDANOV in 1972 does make some provision for surface roughness. It is based on a numerical integration of Fick's diffusion equation, using the universal vertical profiles from the theory of Rossby-similarity in the planetary boundary layer for the wind speed in the form of the velocity deficit and for the turbulent diffusion coefficients k_y and k_z (which may also show empirical x-dependence). This would mean that the resulting concentration distribution - when it is correctly non-dimensionalised - in the lee of a continuous source point, would have to be a universal distribution, i.e. one which is independent of "external" parameters and affected only by the internal parameter for thermal stratification (μ = H/L = internal scale height of the planetary boundary layer over Monin-Obukhov's stability length). But the equation to be solved also contains a term with $Z_o = z_o/H$, which would make the solution dependent on this special external parameter.

It is unfortunate that no universal concentration distribution can be obtained to correspond to each thermal stratification μ; on the other hand it is just this very drawback which makes it possible to take account of z_o.

The corresponding diffusion parameters σ_y (x) and σ_z (x) would therefore have to be evaluated from the solutions $\bar{s}$ $(x,y,z;\mu,z_o,h)$. ("Corresponding" here refers to those parameters which correspond to the assumed bi-normal distribution). Such an analysis should also give the dependence on z_o.

The dependence on z_o - one would guess - probably decreases with increasing source distance x, for example on transition into the meso-scale, and eventually becomes insignificant. But this will have to be tested and proved, and it should be possible to do so using the method described above.

5.2. Considering the vertical profile of wind and exchange coefficients

According to assumption (d) in Section 3.2. the meteorological parameters, i.e. $\bar{u}$, σ_y (x) and σ_z (x) have to be constant with height for bi-normal concentration distribution to be applied. To allow for this, a vertical mean value $\hat{\bar{u}}$ for wind speed is used, and values of σ_y and σ_z are obtained from experiments in a way which assumes that they are in fact constant with height.

With the method of WIPPERMANN and YORDANOV (1972) mentioned in Section 5.1.2., it is possible to get a concentration distribution which allows for the actual wind profile, including the variation of wind direction with height. As Rossby-similarity is used for this there is only one (i.e. the universal) non-dimensional wind profile for each thermal stratification μ (see Section 5.1.2.). The same applies to the two correctly non-dimensionalised coefficients k_y and k_z of turbulent diffusion.

The concentration distribution obtained with profiles of this kind by integration of Fick's diffusion equation should be more realistic than those defined by the usual formulae used to estimate diffusion, which assume a bi-normal distribution.

For each of the concentration distributions computed with these profiles, the corresponding values of σ_y (x) and σ_z (x) would have to be calculated and compared with those given in the usual diagrams (e.g. PASQUILL 1961).

No doubt interesting dependences on thermal stratification, expressed in terms of the internal stratification parameters μ on the one hand, and diffusion categories A - F on the other, will be revealed.

The author would like to take this opportunity to repeat his opinion (WIPPERMANN 1973) that the internal parameter μ for thermal stratification is just the measure for diffusion conditions as depending on the stratification one is looking for; the empirical diffusion categories could probably be replaced by this parameter.

5.3. Considering the unsteady-state aspects

The further one goes into the mesoscale, and the larger the distance from the source, the longer the travel time of the particles becomes, and it is thus increasingly probable that weather conditions will change when diffusion takes place. This may, on one hand, involve changes in thermal stratification caused by diurnal variation of radiation; on the other hand, "synoptic" changes may result from the migration of large-scale eddies (high and low pressure systems), by which mainly the direction of the mean wind changes.

These effects can be eliminated by a method of trajectory calculation (even with large-area grids) as described in Section 4.5. Successive recalculation of such trajectories is carried out in ten-minute steps, which means that a curved trajectory changing with time could be calculated even within a larger grid, and even if conditions within a square of the grid could only linearly be interpolated from the values at the four corners at each time step. It would of course be better to use a "nesting" model as this would also take account of the mesoscale topography of the earth's surface.

It would be quite feasible to reach some agreement with the central office of one of the larger weather services (e.g. Offenbach in West Germany or Bracknell in Great Britain), whereby programming of trajectory estimates could be considered as part of the work now under way to devise nesting models. Normally weather services are not interested in computations of this kind, but the author feels that they could easily be convinced of the need for them.

Trajectory forecasts of this kind should be made for a few important, selected starting points (reactor sites !). Such trajectories computed during the weather service's day-to-day operation could be used to produce statistics. These should refer firstly to the distance between the points of two trajectories with the same distance from the source, one of which is computed, for steady-state conditions using the wind direction and wind speed at the source, the other being based on the actual case of non-stationary conditions and obtained from the trajectory computation. This distance, taken of course as an absolute, becomes larger with increasing source distance and is a measure for the error due to non-stability. Alternatively, these statistics could be made to refer to the final points of the trajectories after 12, 24 or 36 hours. With a sufficiently large set of data (for example daily trajectory calculations over three years) more reliable predictions would be obtainable than current climatological predictions based on statistical wind averages at the source point (see Section 4.1.). The final points of the trajectories would, even with a longer travel time, be by no means equally spaced around the source point; an example of this type of computation was given by RHODE (1973), although this was based on a smaller statistical set.

It would also be quite feasible and practicable to set up a system for forecasting trajectories in emergencies, in collaboration with the data processing centre of one of the major weather services. In the case of a reactor accident the data processing centre would be able to compute the trajectory of the "centre of gravity" of the gas cloud for the next 2-3 days, producing its results 1-2 hours after receiving information on the accident, and reporting back to the location of the accident or to a central authority. Using the dilution factors shown in Table 4, maximum concentration values could then be worked out for each point along the trajectory, as long as there were facilities for measuring the concentration 10 km away from the place of the accident. It would then be possible to compute or forecast all the information for the mesoscale which the current state of our knowledge allows.

6. Bibliography

The following list does not include the papers listed in the bibliography in Section 4.2.

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