Meteorological Factors Influencing the Dispersion of Airborne Diseases [and Discussion]

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Meteorological factors influencing the dispersion of airborne diseases

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Many infective agents are transmitted between host bodies through the atmosphere, sometimes over quite considerable distances. Of prime importance to any potential host is whether or not the agents can be carried to it by the wind from a source-host. The paper briefly outlines how air trajectories are determined and indicates the importance and magnitude of crosswind turbulent spreading. The paper also describes the effect of topography and the likely significance of changes in the wind when the emission persists over several days.

Rain is another factor that can be very important in leading to secondary infections. However, its sporadic nature presents considerable difficulties for the modeller, and this is illustrated.

1. The atmospheric boundary layer

The atmospheric boundary layer, or mixing layer as it is sometimes called in pollution studies, is a layer of variable depth under the direct influence of the underlying surface. It is the layer into which the infective agents of disease are injected and carried passively by the wind. The

Table 1. Percentage of ascents showing an inversion (intensity ⩾ 0.5 °C) within the indicated height range

(z_{im} is the median inversion height, i.e. on 50% of all occasions an inversion lies at or below this height.)

time G.M.T.	z_i/m 0–1	1–75	75–300	300–600	600–900	900–1200	percentage with inversions at top of mixing layer	height of inversion/m mean, $\bar{z}_i$	median, z_{im}
00	59.5	9.5	5.8	3.5	2.8	2.2	85	100	0
06	32.9	12.2	16.2	6.1	4.3	2.9	78	200	130
12	2.4	5.6	4.5	8.1	9.7	5.2	50	800	1700
18	34.2	4.6	3.0	2.8	3.5	3.7	58	200	1000
all hours	33.2	8.0	7.6	5.0	4.4	4.0	—	—	—

boundary layer will respond and evolve as the surface fluxes of heat and water vapour change and as the external large-scale synoptic pattern of depressions and anticyclones itself changes. The layer extends upwards from the surface to where all the turbulent flux-divergences resulting from surface action have fallen virtually to zero. Often the layer is capped by a rise in temperature with height, which inhibits vertical motion and hence the carrying of any of the infective agents out of the layer. At other times clouds may be present with their own circulations and evolution, which may result in some losses from the boundary layer into the atmosphere above.

Typically the height of the boundary layer may be from a few hundred metres to 2000 m by day to perhaps only a few tens of metres, up to say 400 m, by night (see table 1, which

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is based on data collected at Cardington from special balloon-borne ascents). Sometimes no capping inversion is evident, and this is the cause of the surprisingly high median value of the boundary layer height z_{im} at midday.

The layer is characterized by rather rapid mixing caused by turbulent eddies, the scales of which vary continuously from about 1 mm up to the whole depth of the layer in the vertical, and to considerably larger scales for the horizontal turbulence. Normally the vertical growth of any cloud of material injected into the boundary layer through turbulent mixing proceeds until the whole depth is uniformly filled. This usually takes anything from 30 min to 3 or 4 h, by which time the cloud may have travelled anywhere between 5 and 50 km. At the same time horizontal eddies will diffuse the cloud acrosswind and alongwind. For a quasi-continuous source of agents the alongwind diffusion is not important. The crosswind width of the plume after 10 km travel is typically about 5 km, and after 50 km about 20 km when sampled over some tens of minutes. The width therefore grows rather more slowly than linearly. Figure 1,

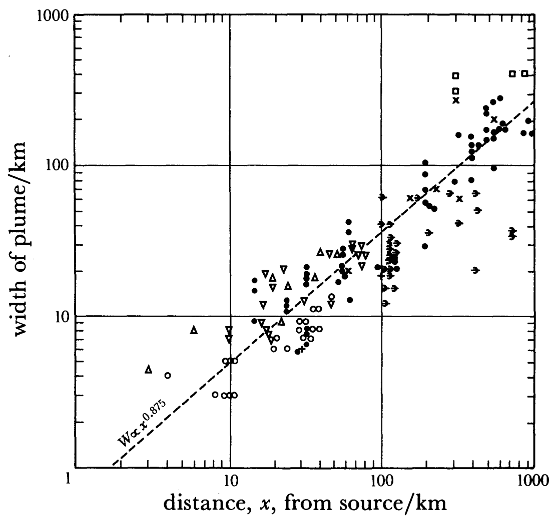

FIGURE 1. Data on the widths of plumes as a function of distance from the source. →, Crabtree (1983) (data collected over the sea); all other symbols, Gifford (1982) (data collected over land).

developed from a figure by Gifford (1982), shows plume-width data collected from a wide variety of sources at ranges extending from a few kilometres to nearly 1000 km. Most of the values (but not all) represent short-period widths, that is widths measured by aircraft flying across the plume, and therefore do not contain the effect of very large eddies that slowly cause the plume to meander. It is interesting to note that the plume widths measured over the sea by Crabtree are generally smaller that those measured over land (presumably because of the relative scarcity of eddies on the scale of the plume width that are most effective in making a plume grow), and the widths grow more slowly with distance beyond a range of about 100 km. For the remainder it seems the width grows like $x^{0.875}$ (where x is the downwind distance from the source). It is interesting that this exponent is identical to that given by O. G. Sutton, a late Fellow of the Royal Society and one-time Director General of the Meteorological Office, in his analysis of the destinations of balloons released during competitions organized for the public at Brighton and in Regents Park, London, in the early 1920s. The original data were collated by Richardson & Proctor (1925), and Sutton's points are included in figure 1.

Theoretically, if the plume gets so large that all the eddies are smaller than the width then the plume grows more slowly at a rate proportional to $x^{\frac{1}{2}}$. No such trend is evident in the land data but may be supported by Crabtree's data over the sea.

2. The advecting wind

The motion of the atmosphere results from spatial variations in pressure. Pressure at any height, being virtually the weight of air in a column of unit horizontal cross section above the height, reflects the vertical density or temperature structure in that column. A column of relatively low density creates low pressure and one of high density, high pressure. Gravity acts on these variations in pressure, resulting in air accelerating from areas of high pressure towards areas of low pressure. However, because the Earth is spinning about its own axis, motions across its surface are subject to an accelerating force at right angles to the motion in obedience to the law of the conservation of total angular momentum. The details of the argument are given in every textbook of basic physical meteorology and will not be repeated here. The force is called the Coriolis force:

$$F = 2\Omega v \sin\phi,$$

where Ω is the angular velocity of the Earth, v is the velocity of the air relative to the Earth's surface, and ϕ is the latitude. In the Northern Hemisphere F is to the right of the velocity vector $\boldsymbol{v}$.

As the air accelerates towards the area of low pressure, the Coriolis force increases and turns the flow to the right until an equilibrium is reached between the pressure force P and the Coriolis force. This happens when the air flows at right angles to P, the pressure-gradient force; that is along the isobars (lines of constant pressure). In this state the velocity is called the geostrophic velocity:

$$v_{\mathrm{g}} = P/(2\Omega \sin\phi).$$

A modified equilibrium is attained if the isobars are curved, which takes into account the centrifugal force as the air tries to follow the isobars, or, if horizontal stresses are important, the stresses originating from the proximity of the rough underlying surface. In this latter case the stresses slow the velocity below its geostrophic magnitude and some cross-isobaric flow towards low pressure is still possible. Thus within the atmospheric boundary layer where stresses are significant we can expect the velocity v to be sub-geostrophic in magnitude (although some important exceptions to this do occur) and backed in direction relative to the isobars. (Backing simply means the direction from which the air is coming is rotated anti-clockwise from the geostrophic direction.) Again this describes the effect in the Northern Hemisphere; opposite trends in direction occur in the Southern Hemisphere. Putting numbers to these effects is often difficult, especially over heterogeneous terrain. Simple as well as quite complex boundary layer models do exist that simulate the observed speed variation and turning of the wind with height over very uniform flat terrain reasonably well, at least in so-called neutral conditions when vertical motions are neither retarded or enhanced by buoyancy forces resulting from changes in potential temperature with height. In unstable conditions, when buoyancy forces tend to increase vertical motions, and in stable conditions when buoyancy forces reduce or inhibit vertical motions, recourse to empiricism is usually necessary, albeit within the framework of Rossby similarity theory, a sophisticated form of dimensional analysis. Details of the theory can be found in many meteorological texts (e.g. Pasquill 1974).

In summary the following simple picture emerges, applicable to rather uniform terrain.

(i) *Neutral stability conditions:* insignificant sensible and latent heat fluxes at the surface. The windspeed increases logarithmically with height. The roughness of the ground is represented by a roughness length z_0, which depends on the physical properties of the surface. Over the sea z_0 is normally about 10^{-4} m, over grassland about 10^{-2} m, over crops about 0.1 m, over typical British mixed agricultural countryside about 0.25 m, and over forests and towns about 1 m. Over about the first 100 metres the wind speed is then given by

$$u = (u_* k) \ln (z/z_0),$$

where z is the height above ground, k is von Karman's constant (≈ 0.4), and u_* is the so-called friction velocity, depending as it does on the magnitude of the surface stress. It is, however, more helpful to relate u_* to the geostrophic velocity v_g. In neutral conditions over homogeneous land $u_* \approx 0.04\, v_g$, whereas over the sea $u_* \approx 0.03\, v_g$. The logarithmic law cannot be expected to hold below about $50 z_0$ because the wakes of individual roughness elements will create local disturbances to the flow.

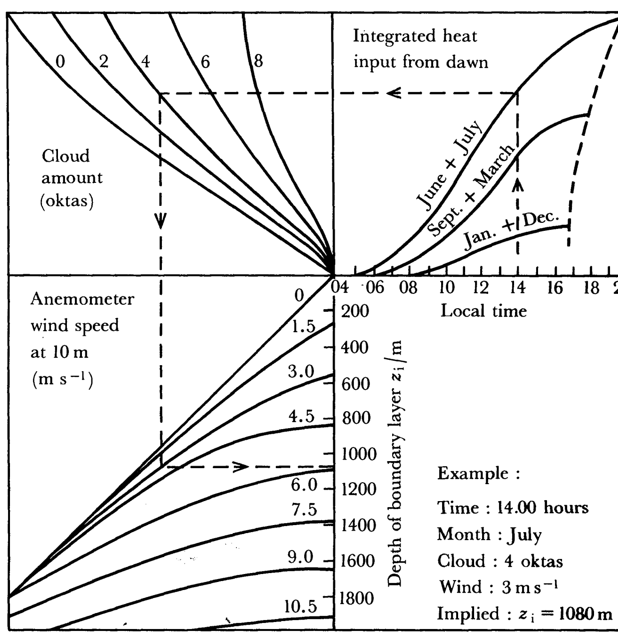

FIGURE 2. A nomogram for estimating the depth of the boundary layer in the absence of marked advective effects or basic changes in weather conditions. The example shows how the diagram is to be used.

Above 100 mm the windspeed, already close to v_g in magnitude, approaches v_g slowly and smoothly until the top of the boundary layer is reached. The height z_i of this top is typically about $0.01\, v_g/f$ in magnitude ($f = 2\Omega \sin\phi \approx 10^{-4}$) so that in mid-latitudes $z_i \approx 100\, v_g$ (z_i in metres, v_g in metres per second).

Wind direction changes almost linearly with height between the surface $z = z_i$ where, in ideal circumstances, the direction is along the isobars. At the surface the wind will be backed through some angle that depend on z_i and z_0 but is typically about 18° over land and 10° over the sea.

(ii) *Unstable conditions* with significant sensible and latent heat fluxes into the air from the surface. These stabilities usually occur only in daytime over land in sunny or partly sunny conditions, or over the sea when the air is cooler than the water. The increased turbulence arising from resulting buoyancy forces tends to increase the depth of the mixing layer. Figure 2 gives a simple nomogram for roughly estimating $h = z_i$ at around 50° N in terms of month, time of day, cloud cover and the 10 m windspeed. Near the ground the wind increases more rapidly at first with height than in neutral conditions but more slowly higher up, whereas the surface-wind backing is less, consistent with the notion that the more vigorously the boundary layer is mixed the more uniform is the wind throughout most of the layer.

(iii) *Stable conditions* are associated with warm air flowing over a cooler surface and a downward heat flux. Potential temperature then increases with height, implying that vertical turbulence is suppressed or partly suppressed by having to do work against gravity. The mixing layer is now often quite shallow and because the downward transport of momentum is depleted by the reduced turbulence the low-level winds are often very light. Because of the poor rate of vertical mixing and the falling temperature of the air near the ground the relative humidity can increase, and mists sometimes form. These conditions may be ideal for the survival of infective agents at high concentrations and the deposition onto vegetation of agents caught on very fine mist droplets.

Rossby similarity theory and Monin–Obukhov similarity theory (another dimensional argument concept appropriate to the lowest surface layer of the boundary layer) are in principle sufficient to determine the nature of the wind profile in both unstable and stable conditions, although in practice both are affected by inhomogeneities in surface conditions and, in stable conditions, by slopes, hills and valleys and other smaller surface features, which can cause substantial deviations from the theoretically predicted state.

Finally, before leaving these particular considerations, it is worth noting that constant-level balloons (tetroons) have been released into the boundary layer to float at about $z = 100$ m and have been tracked out to distances of the order of 200 km by using radar, or in some cases have been returned to the launchers by members of the public with details of their ultimate recovery point. Pack *et al.* (1978) has summarized many of these experiments. As we remember that these are one-level motions and do not fully represent the behaviour of 'free' infective agents, the following are just two of Pack's conclusions:

(i) the backing in southerly flows is on average 20° from the surface geostrophic direction and about 40° for northerly flows;

(ii) optimum fit to the trajectories was obtained by using twice the 10 m windspeed veered by 10° (veering is the opposite of backing). Errors were nevertheless appreciable.

3. Mesoscale motions

Mesoscale motions are not readily quantified either in detail or statistically because they depend on the nature and the state of the underlying terrain, factors which themselves are not easily quantified, nor are the motions clearly distinguished by meteorological observing stations, which are too widely spaced for this purpose. The scale of these motions lie in the range 1–100 km. In recent years field experiments have begun to unravel their nature. Satellite pictures are also proving very useful in the study of some kinds of mesoscale motions and have revealed occasional coherence over much larger scales, e.g. lines of cumulus clouds, indicating the existence of large boundary layer rolls maintaining the clouds, extending many hundreds of kilometres.

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Other examples of mesoscale circulations are sea and land breezes, flows associated with large convective clouds, urban circulations and topographically induced flows.

Sea breezes are quite common in summer around coasts, although in the U.K. they are usually very weak affairs. An appreciable breeze only occurs when air temperatures over land exceed temperatures over the adjoining sea by more than 2 °C, when the 900 m wind is less than about 7 m s^{-1} and has an offshore component, particularly if it comes from the left as one faces inland (D. Houghton, private communication). Sea breezes with onshore gradient winds are relatively uncommon for dynamical reasons. Once established the inflow bottom layer (sea to land) is typically about 500 m deep and is some one-third to one-quarter of the total atmospheric boundary layer depth, the remainder being the outflow region. The breeze penetrates further and further inland during the day, reaching 20–100 km in ideal conditions in the U.K. The circulation extends some 10–15 km out to sea, and speeds of 3–5 m s^{-1} in the inflow region are quite normal. In light geostrophic winds the direction of the inflow starts by being at right angles to the coast, but as the breeze develops it often turns to the right and flows some 20–30° to the coastline.

Land breezes at night form in response to the nocturnal cooling of the land but are generally more feeble and rarely exceed about 60 m in depth. Even a light onshore component of the geostrophic wind will often completely suppress it. If, however, high ground exists only a few kilometres inland, downslope flows may reinforce the land breeze provided that the pressure gradient is slack and the skies are virtually cloudless.

Large storms have a very dramatic effect on the boundary layer and its contents. They represent a very real breakdown of the normal structure. Infective agents could be drawn into the cloud and virtually lost without necessarily thereby reducing ground-level concentrations, only their horizontal extent. Associated rain, however, might deposit large numbers of agents on to the ground. Thus the actual net effect of such storms as far as secondary infections are concerned is very uncertain; all that can be said is that they can represent regions of high risk.

The effect of hilly terrain on airflow and the dispersion of diseases is complex, and is the subject for continuing research employing complex theory, wind-tunnel modelling and major field experiments. The subject is much too large and difficult to discuss adequately here; it will suffice to mention just a few qualitative points of importance to the subject.

(i) Mountains are associated with increased condensation of water vapour in the air, increased rainfall and the wet removal of airborne vectors.

(ii) In near-neutral conditions, simple rounded hills cause increases in windspeed and turbulent velocities at the crest by a factor of the order of $2h/l$ (where h is the hill height and l is the half width of the hill half way up). Rougher hills also increase the velocities of course, but the factor is much harder to predict.

(iii) In stable conditions, air has to do work if it rises, meaning that kinetic energy has to be lost to potential energy. If the windspeed is too low or the ridges too high the air can only be trapped, having to find its way around the obstacle whenever this is possible. This is the basis of the model referred to by Gloster (this symposium).

(iv) In stably stratified airflows areas of high ground can, for related reasons, generate persistent *horizontal* lee eddies on the same scale that can significantly distort the path of any infectious agents. An example of this is shown in figure 3 in which a large sub-synoptic scale eddy downward from the North Yorkshire Moors is evident from aircraft-measured winds within the boundary layer. The eddy has been detected on three separate occasions during recent pollution sampling flights off the English coast.

For fuller details on the effect of hills on airflows and dispersion, reference should be made to Hunt & Mulhearn (1973), Mason & Sykes (1979, 1981) and Puttock & Hunt (1979).

Several theoretical models have been and are being developed to study these and other mesoscale motions. In general they require considerable computational effort and a very detailed database, and are therefore more suited to the study of special events rather than regular operational use. Details of two such models can be found in Carpenter (1979) and Seaman & Anthes (1981).

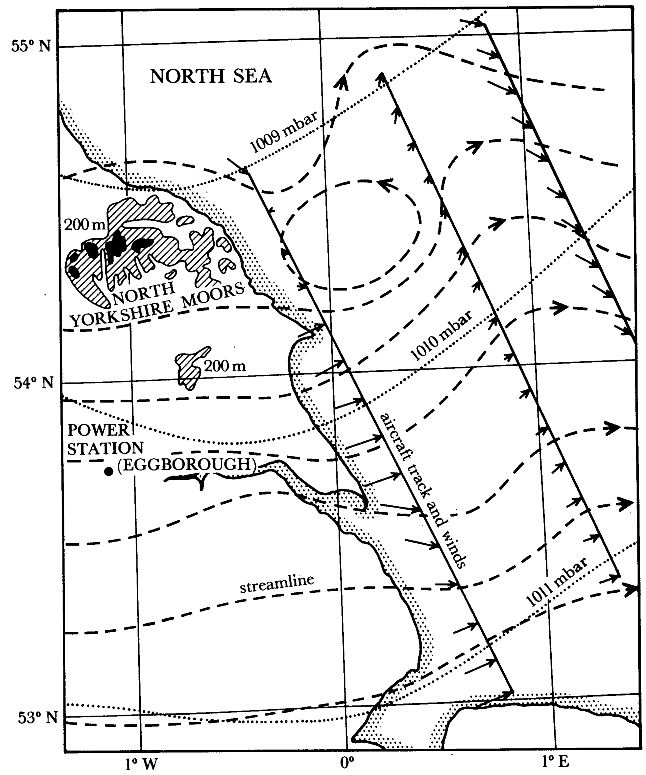

FIGURE 3. An apparent mesoscale horizontal eddy downwind from the North Yorkshire Moors implied by winds measured by the Meteorological Office's Hercules aircraft on 18 June 1980 during an experiment to study the plume from Eggborough Power Station (Crabtree 1983).

4. Errors in wind estimates and synoptic-swinging

When a source of infection has been located and the area of possible subsequent secondary infection is being sought, an appropriate wind field is needed. The winds may be estimated in a variety of ways, all of which involve errors to a greater or lesser degree. If the release of infective agents into the air takes place over a very short time, the impact of these errors will be at a maximum. They arise from a variety of causes. Sometimes winds are estimated by interpolating between meteorological-station winds. These winds are obtained by using robust instruments built to last in all sorts of weather and which are subject to error, especially at low windspeeds when concentrations of agents may be at their highest and the risks of secondary infection are appreciable. The anemometers may be subject to siting problems and could be influenced in certain wind directions by the wakes from nearby woods or buildings, or by distant topography. Interpolation will ignore important static or transient mesoscale motions on a scale less than the separation of the stations. The same would be true or partly true of winds obtained from numerical models by using both pressure and wind observations, particularly if the grid size is more than just a few kilometres. Over typically heterogeneous terrain it seems from experiments that individual trajectories determined by rather simple means can be out by as much as 20 or 30 km after just the first 100 km travel.

Normally, however, the release will persist over a day or several days, during which time the wind direction will probably have changed quite considerably. The downwind area affected by the agents will then be dominated by these changes, swamping the errors in each trajectory determination, except perhaps at short range where the effects of marked local topography may persist and are inadequately allowed for in the wind-flow determinations. At longer range even these effects are likely to be rather unimportant. To give some indication of the relative importance of the wind direction changes compared with other effects already discussed, I shall refer to a study by Smith (1980), who considered some 2100 sequential trajectories arriving at a point in southern Norway, calculated by using surface geostrophic winds, and evaluated every 12 h. The trajectories show a fairly systematic swing from one to the next, although on many occasions the swing is first one way and then the other. The changes reflect the general movement of synoptic-scale features (depressions and anticyclones), and for this reason the process has been called 'synoptic-swinging'. The bearings, θ, where the trajectories crossed arcs of radii 250, 500, 750, 1000, 1250 and 1500 km were extracted. Probability distributions were constructed of swing for given sampling times and ranges, and from which it was possible to construct the curves shown in figure 4. For a sampling time (or release time of the infectious agent) of 30 h and at a range of 100 km for example, it can be deduced that the mean swing will be less than 15° on 10% of occasions, less than 52° on 50% of occasions and will only exceed about 140° one time in ten. These values can be obtained numerically using the following approximate formulae:

$$\text{if } \delta\theta_w = 2.2\times10^{-2}\ T^{1.16}\ X^{-0.125}, \quad \text{then} \quad \delta\theta < \delta\theta_w \ 10\,\% \text{ of the time;}$$

$$\text{if } \delta\theta_w = 0.19\ T^{0.85}\ X^{-0.125}, \quad \text{then} \quad \delta\theta < \delta\theta_w \ 50\,\% \text{ of the time;}$$

$$\text{if } \delta\theta_w = 1.1\ T^{0.64}\ X^{-0.125}, \quad \text{then} \quad \delta\theta < \delta\theta_w \ 90\,\% \text{ of the time;}$$

where $\delta\theta$ is in radians, T is in hours and X, the downwind distance, is in metres. On average, trajectories emanating from a source of infection exhibit a tendency to bend very gradually towards the east in response to the overall mean west–east zonal flow in temperate latitudes. The average change of angle, $\Delta\theta$, per 100 km of travel in this sense was found to be closely given by

$$\Delta\theta = 0.25 + 1.3\sin(\theta + 70),$$

where θ is in degrees). This property is illustrated in figure 5, where the number of crossings of arcs of specified radii in 30° sectors is displayed as a direction rose. The larger the range the greater is the tendency for the crossings to be towards the east. However, the effect only becomes appreciable at ranges when in most cases the risk of secondary infection would be negligible.

5. Surface deposition by rain

Many substances emitted into the atmosphere are subject to removal by precipitation either by a rain-out process, where the substances gets absorbed into, or forms the nucleus for, growing rain-drops within the clouds, or by a wash-out process, where the pollutant is captured by the drops as they fall through the mixing layer below cloud, or by both processes.

In considering the effect of rain on airborne diseases the first issue of importance is whether or not the vectors can survive the experience of being brought down to ground in this way. If they can, then rainfall can be a very efficient way of infecting a new area, if not then it can be very important in reducing the risk of secondary infections. Either way we must know where rain has occurred along the plume. At present rainfall is usually assessed from observations made at a rather sparse network of meteorological stations. The situation is particularly acute over the sea, where observations are either non-existent or are few in number and quantitatively

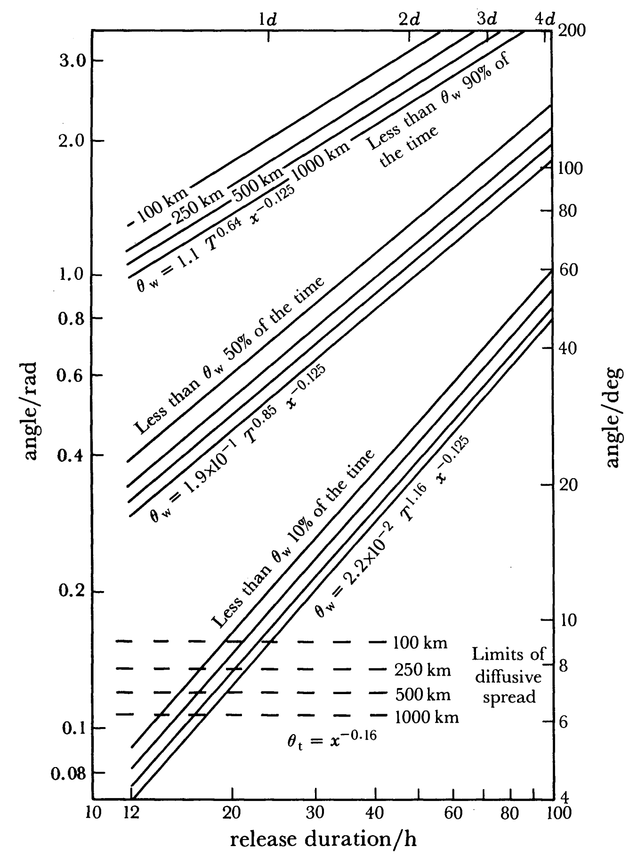

FIGURE 4. Plume width as a function of distance from the source and release duration for different probabilities.

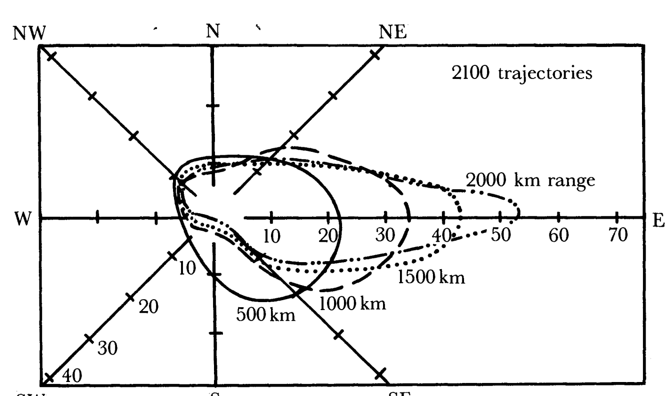

FIGURE 5. The property shown in figure 4 is illustrated here, where the number of crossings of arcs of specified radii in 30° sectors by 2100 forward trajectories is displayed as a direction rose. The larger the range the greater is the tendency for the crossings to be towards the east.

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rather uncertain. Even over land the locations of the stations are not always without bias: the majority are in open-level terrain, in valleys or at coastal sites. Elevated mountainous sites, where the rainfall is often much greater than elsewhere, are for obvious reasons few and far between. The rainfall affecting the trajectory has to be deduced either from spot observations made every 1 or 3 h at these stations of from accumulated-rainfall observations made every 3 or 6 h.

Ultimately this situation may improve as more and more areas and surrounding seas are covered by weather-radar networks (as are large parts of England and Wales), which can make and record quantitative measurements of rainfall intensity with good spatial and temporal

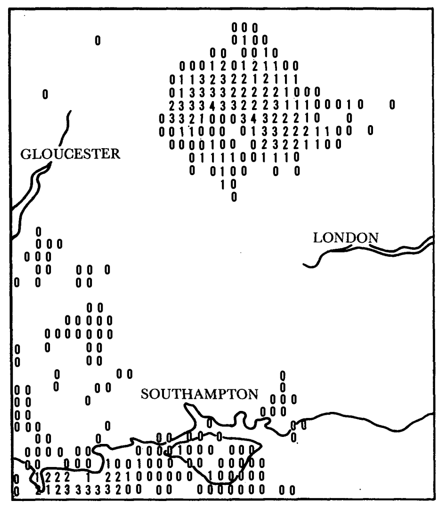

FIGURE 6. Part of the weather-radar display at 02h00 G.M.T. on 1 May 1980. For purposes of display, rainfall intensities are classified as follows: 0, 0–1 mm h^{-1}; 1, 1–2 mm h^{-1}; 2, 2–4 mm h^{-1}; 3, 4–8 mm h^{-1}; 4, 8–16 mm h^{-1}; 5, 16–32 mm h^{-1}.

resolution (the U.K. radar assesses intensity over 5 km × 5 km squares every 15 min). It is clear that such data are potentially invaluable and should lead to significant improvements in this particular respect in the future. However, they may not provide all the answers, especially in mountainous areas The radar cannot see into areas in the shadow of intervening mountains, and rainfall intensity is not constant with height but tends to increase significantly in the lower layers out of view of the radar in mountainous areas. The reason for this is the effect of so-called 'feeder clouds' forming in the lowest layers of the air as that air is forced up over high topography. These clouds add considerable amounts of water to the drops falling through them from above.

Another problem with rainfall and the removal of infective agents is that many cases of really very heavy rainfall are associated with rather complicated baroclinic flow fields when the simple trajectory-determination techniques touched on earlier may prove to be unreliable.

Finally, figure 6 shows an example of the output from the U.K. weather-radar network, run jointly by the Meteorological Office and the Royal Signal and Radar Establishment at Malvern, and clearly illustrates the sporadic nature of rainfall in some situations. On this day, 1 May 1980, all four radars available at that time were in operation when a series of active rain-cells spread across central southern England and south Wales from the east-southeast between midnight and 07h00 G.M.T. Individual cells, typically 100–150 km across, could be followed across the area for most of this period and smaller 'hot-spots' of heavy rain associated with individual cloud-cells embedded in the larger rain-cells persisted for around 30–45 min, being picked up on two or three consecutive displays given at 15 min intervals.

Examination of cases like this shows that the standard technique involving interpolation between reports of rain from meteorological stations would give a totally false impression of the rain experienced along any one plume, and the likely deposition of agents, because the interpolation would inevitably smooth the rainfall pattern in both time and space.

6. Conclusions

Determining the history and fate of airborne infective agents is difficult, subject to error and demanding of the most sophisticated techniques of the meteorologist. The paper has briefly outlined many of these difficulties and has pointed to some developments in methods that may give better information in the future.

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Discussion

R. S. Scorer (*Imperial College, London, U.K.*). The picture of a plume from a source undergoing gradual dilution by spreading upwards and sideways is too simple and may be quite misleading as to the effects. Volumes of pollutant are drawn out so that starting as a compact mass they are converted into a tortuous sheet on both sides of which the gradient of pollution concentration is large. This is less important when we are considering the spread of a pollutant than when considering its detection by sense organs. Thus the human nose can detect a few molecules of a mercaptan from an oil refinery plume on first entering the edge of the plume, but may be uncertain as to the presence of the plume in its very centre where the concentration is high but the pollution fairly continuous. Thus the nose detects a rapidly rising concentration especially from zero or very low concentration. Moths may likewise be more sensitive to large gradients of concentration of a scent than to a steady but higher concentration. This is relevant to the accuracy of a picture of the mechanism of dispersion. Molecular diffusion operates on

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the small scale of centimetres only, after several minutes; eddy diffusion consists of drawing the pollutant out into highly contorted sheets with a continual intensification of the gradients by the stretching due to the shearing motion.

Dr Smith mentioned the street (he called it a vortex) stretching many miles downwind of Ailsa Craig. Satellite pictures often show single streets extending hundreds of miles across the sea. I have seen a single street, which might imply rather little lateral spread, for 1000 km downwind of Madeira, which could carry insects to, say, a Canary Island, if it happened to be suitably placed, and in fairly high concentration considering the distance.

Similar solitary streets are often seen over the sea and could carry a particulate load from Denmark to England, reaching the east coast as a fairly narrow plume.

F. B. Smith. Professor Scorer firstly describes the well known phenomenon of dispersion by turbulent eddies in a correct and vivid fashion. I have no disagreement with this. However, as applied to the context of my talk and to this Meeting, my reply is similar in some respects to my reply to Dr Green after the next paper. A potential host is only likely to become infected if its natural body defences are swamped by incoming infection, and this implies a persistent exposure over some time when the fine detail of the virus plume structure (referred to by Professor Scorer) is largely smoothed out by time-averaging as the plume material sweeps by.

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