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THORNEY ISLAND DATA AND DISPERSION MODELLING

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Summary

The Thorney Island experiments have produced a large body of data suitable for the validation of mathematical models of dense gas dispersion. In this paper we consider the process of comparison of model predictions with such data.

First, we consider what features of dense gas dispersion models are tested by various types of field experiment, and the new information provided by the Thorney Island trials.

The analysis of ambient conditions, often imperfectly steady and uniform, is discussed, along with an assessment of random and systematic measurement errors. The characterisation of turbulence and atmospheric stability takes a large part of this analysis. Before a comparison is made with data, it is desirable to know exactly what any model is intended to predict (usually some sort of average), and, in a final assessment, how much one experiment might be expected to deviate from the "average".

1. Large-scale experiments for the validation of dense gas dispersion models

Dense gas dispersion experiments can be performed for a number of reasons. Most straightforward, perhaps, is the direct simulation of an accidental release. Alternatively, the purpose may be to validate wind tunnel or water flume techniques. Or the data can be intended for the validation of mathematical models. It is this last possibility which we take to be the main aim of the Phase I Thorney Island trials. (Included in this is the investigation of the physical processes of dispersion, on which the models should be based.)

The process of model validation ideally involves much more than the measurement of one parameter, for example the distance to a lower flammable limit, and comparison with the corresponding model prediction. It is important that many aspects of cloud dispersion be measured and comparisons made with predictions in each area. Hence there is a need for a variety of instruments and numerous measurements in such experiments. Then the

mechanisms included in the model can be assessed individually, removing the danger of producing a correct overall answer as a result of two errors cancelling one another.

It is unlikely that every important mechanism can have a dominant effect in one type of experiment. So there is also a need for a variety of field experiments to be performed. An initially high cloud of gas might behave quite differently from gas which evaporates from a liquid pool; the first cloud has more gravitational potential energy, much of which is transformed during slumping to kinetic energy of the large-scale flow or of turbulence. Both kinds of experiment are relevant to possible incidents, in one case sudden releases of liquids stored under pressure well above their boiling points [1, 2], in the other, spills of refrigerated liquids.

The use of ambient temperature gas eliminates confusing thermal effects when one is simply trying to understand dense gas dispersion. Releases of freon, for example, allow models to be checked for this basic case. But then experiments with a gas such as methane (LNG) are also needed to show the effects of thermally driven motions and changing buoyancy on the dispersion process. An intermediate case with liquid nitrogen for instance, which is cold but where the gas cannot become lighter than air, might also be interesting.

Steady-state (continuous) release experiments are useful in that they model possible incidents such as steady leaks from pipes and vessels. They also represent one extreme of the spill types with which models should be able to cope; and they allow the possibility, in principle, of studying short-term concentration fluctuations in one experiment. However, an assessment is also needed of models' ability to deal with the strongly time-dependent behaviour exhibited by instantaneous or short-time releases; so such experiments are also performed. Furthermore, in practice, much stronger dense gas effects can be achieved in instantaneous spills. The variety of experimental options and their importance has been discussed at greater length by Puttock et al. [3].

The Thorney Island trials have involved instantaneous releases of high clouds of gas at ambient temperature. Prior to these, large-scale field experiments have generally involved spills of liquefied gases [3]. We shall mention the two most recent tests of this kind. The 1980 "Burro" series of tests performed by the Lawrence Livermore Laboratory at China Lake [4] comprised eight releases of forty cubic metres of LNG. Each spill lasted about three minutes. At Maplin Sands, Shell Research performed twenty-four steady-state releases of refrigerated liquefied propane and LNG onto the sea [5]. In addition there were a few instantaneous releases of liquid. For propane, thermal effects are not very important in dispersion and so these spills are similar to releases of ambient temperature dense gas. The contrast with the LNG spills has shown the importance of thermal effects in the case of LNG [6].

Instantaneous spills of liquid do not produce instantaneous clouds of gas

because of the time needed for evaporation. Only for low winds is the gas release effectively instantaneous. Thus China Lake and Maplin each only produced one instrumented spill which could be classified as "instantaneous gas" [3]. The fifteen unobstructed dense gas spills at Thorney Island add considerably to the database in this area.

The other main contrast between the Thorney Island trials and liquefied gas spills is the initial height of the cloud. To what extent does the drop from a height of thirteen metres affect the dispersion when compared with a

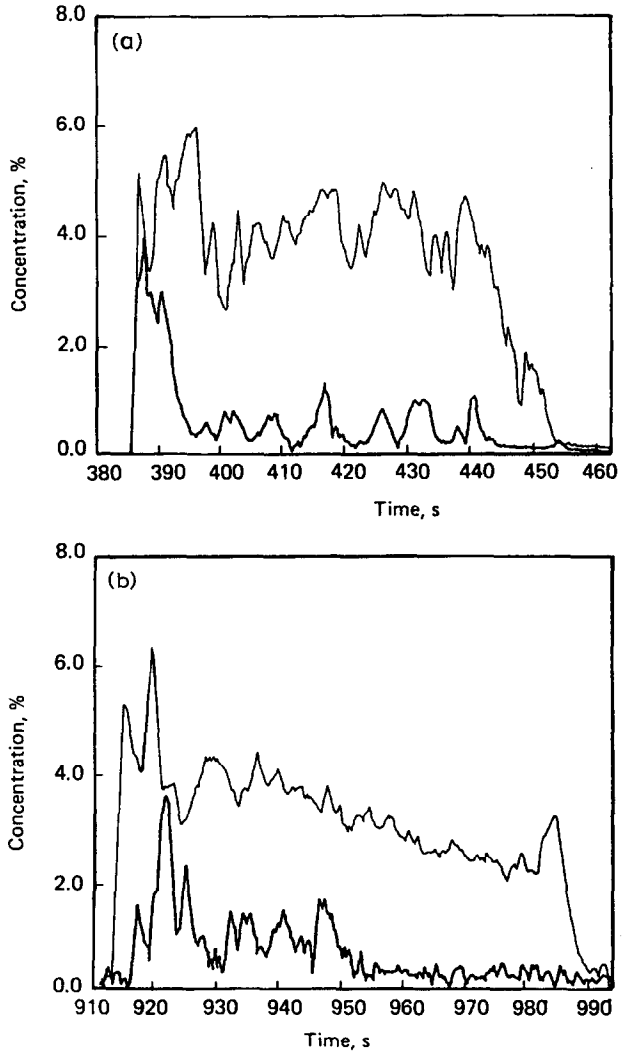

Fig. 1. (a) Gas concentration measurements in Thorney Island Trial 7, taken at 100 m from the spill point, 45° to the wind direction, and 0.4 m and 2.4 m above the surface. (b) Gas concentration measurements in Maplin Sands Trial 63, taken at 111 m from the spill point, 59° to the wind direction, and 0.6 m and 2.3 m above the surface.

cloud which starts and remains low? In this respect comparison of, for example, Thorney spill 7 and Maplin spill 63 will be interesting. As can be seen from Table 1, conditions for these two spills were very similar, the main differences being the volume of gas used and the surface roughness. Examples of concentration measurements from similar sensor locations are shown in Fig. 1.

TABLE 1

Similarity of conditions for Thorney Island Trial 7 and Maplin Sands Trial 63

	Thorney Island Trial 7	Maplin Sands Trial 63
Initial volume of gas	2000 m^3	4000 m^3
Relative density	1.75	1.86
Wind speed	3.2 m/s	3.4 m/s
Stability class	E	D

2. Analysis of the meteorological data

Proper interpretation of observed dispersion behaviour requires careful and detailed analysis of meteorological data. In contrast to the laboratory, the flow field in the atmosphere is generally not stationary and homogeneous. It has to be established first whether variations in time and/or space of "mean" wind speed and direction are significant during the data recording period for a trial. If so, the determination of a meaningful value for mean wind speed and direction is not straightforward. Then the averaging period for a particular sensor reading should coincide with the period when the cloud is drifting past that sensor. (Close to the source the wind field may be considerably influenced by the cloud's presence. Instrument readings affected in this way obviously should not be used in the characterisation of the ambient flow field.)

Visual examination of the recordings is required to determine if the wind field was stationary during the time the cloud was passing. Equally, if mean values calculated in this way differ significantly between sensors, one must conclude that the part of the wind field affecting the cloud was non-homogeneous and/or non-stationary. Comparison of the dispersion behaviour with results of any existing model will be very dubious for such environmental conditions.

Careful determination of the atmospheric stability takes a large part of the effort in the analysis of the micro-meteorological data. Close to the source, when gravitational spreading velocities are not small compared with the wind speed, the dispersion will hardly be influenced by the turbulence characteristics of the wind field. Moving downwind the dispersion will be more and

more determined by the wind field properties. Very toxic gases have to be diluted to concentrations in the ppm range to be non-hazardous. For the dispersion of these gases the atmospheric stability will be an important parameter.

The lower limit of resolution of the Thorney Island gas concentration measurements was 0.1%. This corresponds to $\Delta\rho/\rho = 0.1\%$ (where $\Delta\rho$ is the density difference between the gas/air mixture and air), for an initial relative density of 2. For Pasquill class F we find close to the earth's surface a temperature gradient of some $-0.5°C$ per 10 m, corresponding to $\Delta\rho/\rho = 0.17\%$ per 10 m. If we assume the cloud to be 10 m high at the downwind position where the 0.1% concentration is found, the density gradient due to the presence of the cloud amounts to 0.1% per 10 m. So only in very stable or unstable situations the dispersion will be influenced significantly by the atmospheric stability, and then only in the far field, where the concentrations are lower than, say, 1%.

2.1 Turbulence classification schemes

Several turbulence classification schemes are used to determine the Pasquill stability classes [7]. They are specifically designed for use in cases in which there is only a very limited number of routine measurements or observations available on which the atmospheric stability estimate can be based. Often these methods give no consistent results [7], as is also found for the Thorney Island trials, where application of the different schemes produced a wide variety in stability classes [8]. For the well-instrumented Thorney Island experiments, methods are available to characterise the atmospheric stability, e.g. in terms of the Monin—Obukhov length, L, which have a more direct relationship with dispersion than, say, observations of cloud cover. Since the dispersion of the clouds was monitored only over a relatively small distance, the vertical growth of the clouds will have been restricted over that distance to the height of the atmospheric surface layer, where the well-proven Monin—Obukhov similarity theory has found wide application. This theory also gives good results in eddy diffusivity models for calculation of the vertical dispersion of passive contaminants [9]. Hence it seems obvious to use the surface layer parameters friction velocity, Monin—Obukhov length and roughness length for characterisation of the turbulent wind field. Some heavy gas dispersion models need a stability classification in terms of the Pasquill classes; using the graph of Golder [10] these classes can be related to the Monin—Obukhov length, L, for given roughness length, z_0. Several methods exist to determine z_0, L and the friction velocity u_* from the measured meteorological data. Two of them will be discussed below: profile fitting and direct flux measurements. At least two methods should be used to allow for a check on consistency.

Crosswind dispersion of passive contaminants, characterised by σ_y as a function of downwind distance x, has been found to correlate well with the measured standard deviation of wind direction fluctuations σ_θ [11]. This

correlation is well represented over the range from 0 to 1 km by

$$\sigma_y = 0.028\,\sigma_\theta x^{0.85},$$

where σ_y and x are in metres and σ_θ in degrees.

Care should be taken in selecting the averaging period for the determination of σ_θ since, as a result of large-scale motions in the atmosphere, σ_θ tends to increase with increasing averaging time. These large-scale motions, however, do not contribute to the dispersion but cause the path of the cloud to meander.

For the same reason sudden changes in wind direction should be excluded when calculating σ_θ. The recorded time series have to be examined visually to detect such changes. The sonic anemometer readings have to be used for the calculation of σ_θ since the lower limit of resolution of the wind vanes is too large for this purpose.

2.2 Computation of the surface layer parameters from measured wind and temperature profiles

The roughness length z_0 of the site can be determined from measurements during high wind speed episodes, when the turbulent properties of the wind field are not influenced by buoyancy effects. The wind velocity profile is then properly described by

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

where k is the von Karman constant. The roughness length can be determined by fitting this profile to the measured wind velocities at different heights, z; extrapolation of the fitted logarithmic curve to the point where $\bar{u} = 0$ gives the value of z_0. This method requires extreme accuracy of the cup anemometer readings; this is why careful calibration and regular checking of the output of these instruments through the data collection system was absolutely necessary.

The best way to determine u_* and L from the wind and temperature profile is to fit both profiles simultaneously to the measured profile data. This method has been successfully applied by Nieuwstadt [12], who used the Kansas profile relations. Extreme accuracy of the temperature profile measurements is required, ideally to some 0.02°C.

The platinum resistance thermometers mounted in Marex screens as used for the profile measurements during the Thorney Island trials have an accuracy of only 0.3°C according to the manufacturer's specification. However, by careful individual calibration this accuracy can be improved probably to better than 0.1°C. Regular checking of the readings of these instruments through the data collection system was also necessary to make them useful for the purpose of profile fit applications.

Measured profile data could be obtained during the trials from 5 cup anemometers and 5 thermometers, to produce with a profile fit method the

surface layer turbulence parameters u_* and L. Obviously such a method is more accurate than turbulence classification schemes which use only measured data at two levels, like the Richardson number method [7]. However, the profile estimates may be misleading since a constant flux layer may not have been re-established after the sea—land transition. There is thus all the more reason to use a second method to determine surface layer parameters.

2.3 Determination of surface layer parameters from flux measurements

Since $u_*^2 \equiv \overline{u'w'}$, the friction velocity u_* can directly be determined from the covariance of the fluctuating parts of the wind speed components in the vertical (w') and mean wind direction (u').

For high wind, neutrally stable conditions the wind profile is described by $\bar{u}(z) = (u_*/k) \ln(z/z_0)$. From this expression z_0 can be calculated, using measured values of $\overline{u'w'}$ and $u(z)$. The Monin—Obukhov length is defined by [7]:

$$L \equiv -\frac{u_*^3 \bar{T}}{kg(\overline{w'T'} + 0.61 \bar{T} \overline{w'q'})}$$

where T is the absolute temperature, g the gravitational acceleration and q the specific humidity (in kg water per kg moist air). Above land the term $0.61 \bar{T} \overline{w'q'}$ is generally negligible compared with $\overline{w'T'}$. For the Maplin Sands site we found that both terms were of the same order and since the Thorney Island site is close to the coast it cannot a priori be assumed that the humidity flux term can be neglected.

Meaningful covariance measurements in the atmosphere are far from easy to obtain. During a data collection period, lasting typically for 1000 s in the Thorney Island trials, strong variations in wind speed or temperature with large time scales may occur. These large-scale periodicities or trends in the signals are considered not to belong to the surface layer turbulence and make the signals statistically non-stationary. Since the Reynolds decomposition is only meaningful for stationary signals we have to remove large-scale variations from the time series. For the analysis of the Maplin Sands data we smoothed the time series using a moving average procedure, the smoothed signals were then subtracted from the original time series to obtain the fluctuating components. The averaging time of the moving average procedure should be some two orders of magnitude larger than the integral time scale of the surface layer turbulence.

Ideally, averaging times in the calculation of turbulent fluxes in a stationary mean flow should approach infinity. Since the data collection periods are rather small in the Thorney Island trials, it is important to have a feeling for the random errors that occur due to the rather short averaging times. A discussion of these errors can be found in references 14 and 15, based on the expression for averaging errors, given by Lumley and Panofsky [16].

For the covariance of two signals f_1 and f_2 this expression for the relative

error ϵ reads:

$$\epsilon^2 = \frac{2\tau_i}{t_a}\left[\frac{\overline{(f_1'f_2')^2}}{\overline{(f_1'f_2')}^2} - 1\right],$$

where τ_i is the integral time scale of $f_1'f_2'$ and t_a is the length of the averaging time interval. Sreenivasan et al. [15] obtained integral time scale values from measurements in the surface layer over the ocean for $z/L = -0.05$. He found $\tau_i\bar{u}/z = 1.2$ for $u'w'$ and $w'T'$ and $\tau_i\bar{u}/z = 1.1$ for $u'q'$. The value of the term

$$\left[\frac{\overline{(f_1'f_2')^2}}{\overline{(f_1'f_2')}^2} - 1\right]$$

as a function of z/L as determined from the Kansas data is given by Wyngaard [14] for $u'w'$ and $u'T'$ and reproduced in Fig. 2.

Five of the ten sonic anemometers at the Thorney Island site were deployed at $z = 2$ m; at this level the values of z/L will always be small and we find from Fig. 2 that

$$\left[\frac{\overline{(f_1'f_2')^2}}{\overline{(f_1'f_2')}^2} - 1\right] \approx 10.$$

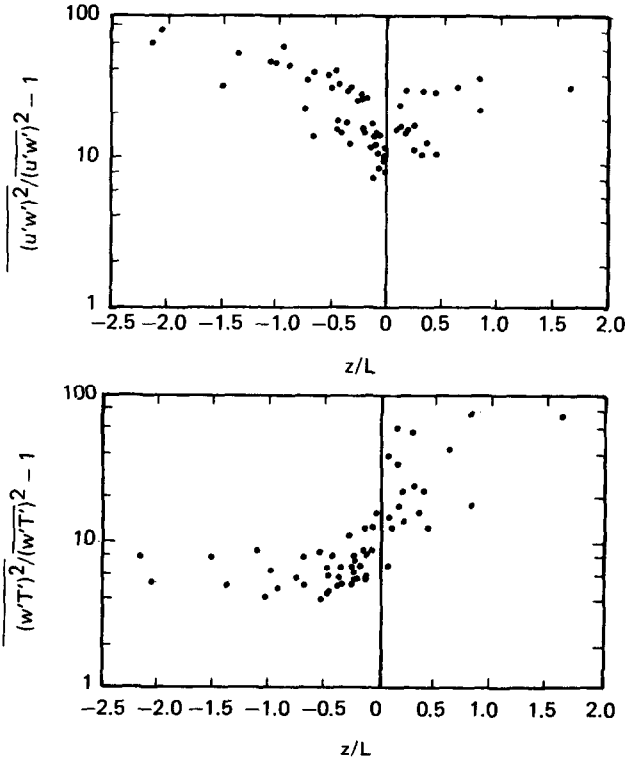

Fig. 2. The variances, about the mean, of stress (upper) and heat flux (lower) and their stability dependence as determined from the Kansas data.

Sreenivasan found as values for this group: 15 for $u'w'$, 32 for $u'T'$ and 22 for $u'q'$. For the Thorney Island trials a value of 10 is most appropriate since this was derived from measurements over land. We then arrive at a relative error estimate, given by

$$\epsilon \approx \left(\frac{20\,z}{\bar{u}\,t_a}\right)^{1/2}.$$

For a low wind speed of 2 m/s, a t_a equal to a typical data series length of 1000 s for the Thorney Island trials, and $z = 2$ m we find $\epsilon \approx 15\%$ which is quite acceptable. This random error is of course further reduced when an average flux value from all anemometers is used. The errors can become much larger when data from sonic anemometers deployed at higher levels than $z = 2$ m are used. On the other hand, the error will decrease with increasing wind speed. To keep ϵ as small as possible the averaging time for surface layer flux measurements should be taken as large as the recorded time series permit and should not be confined to the period of time needed for the cloud to drift past a sensor location.

As well as random errors in the determination of covariance there are also some important systematic errors in these measurements. We will first deal with the influence of instrument tilt and flow distortion on the measured values of $u'w'$ which we will denote by $\overline{u'_m w'_m}$. This error has received attention recently from several authors [17—19], who derived expressions to correct for the errors introduced by these effects. Instrument tilt as well as flow distortion by supporting structures will result in a non-zero value of $\overline{w_m}$, which should be zero over a flat site.

The tilt equation [17] is

$$\overline{u'w'} = \overline{u'_m w'_m}\cos 2\alpha + \tfrac{1}{2}\sin 2\alpha\,(\overline{u'^2_m} - \overline{w'^2_m}),$$

where α is found from $\tan\alpha = -(\bar{w}_m/\bar{u}_m)$. If the velocities are measured far from the flow distorting structure and/or approximately above or below this structure the flow distortion equation derived by Dyer [17] reads

$$\overline{u'w'} = \overline{u'_m w'_m} + \overline{u'^2_m}\tan\alpha$$

which is very similar to the tilt equation for small values of α, apart from the absence of $\overline{w'^2_m}$ in the second term. Dyer [19] stated that Wyngaard's more elaborate theory [18] suggests that a flow distortion equation would be similar to the tilt equation but with the $(\overline{u'^2_m} - \overline{w'^2_m})$ term replaced by $(\overline{u'^2_m} + \overline{w'^2_m})$. However, in practical situations it will be difficult, if not impossible, to discriminate between both error sources and therefore we suggest the use of the following correction formula:

$$\overline{u'w'} = \overline{u'_m w'_m} - \frac{\bar{w}_m}{\bar{u}_m}\,\overline{u'^2_m},$$

corresponding with Dyer's flow distortion equation. In practice $\overline{u'^2_m}/\overline{u'_m w'_m} = 6$, and the correction amounts to about 10% per degree. This correction turns

out to be substantial but can be easily calculated from the measured wind velocity data.

To conclude this section we will pay attention to some special features of the "heat flux" measurement with sonic anemometers and an important systematic error in this measurement. The sonic temperature measurement is based on the temperature dependence of the sound velocity, c, in air. However, c also depends on the specific humidity, q, and it can be shown [20] that the "temperature" $\bar{T}_m$ obtained from the temperature output of the sonic anemometers is given by $\bar{T}_m = \bar{T}\ (1 + 0.51\ \bar{q})$. The "heat flux", $\overline{w'T'_m}$, measured with the sonic anemometer can be represented by

$$\overline{w'T'_m} = \overline{w'T'} + 0.51\ \bar{T}\ \overline{w'q'} - 2\frac{\bar{T}\bar{u}}{c^2}\ \overline{u'w'}$$

if the temperature is measured along the vertical axis, as most sonic anemometers do [20]. The correction term

$$-2\frac{\bar{T}\bar{u}}{c^2}\ \overline{u'w'}$$

becomes very important at high wind speed since it increases with $\bar{u}^3$. For this reason the sonic temperature is normally not used to determine heat fluxes. But Schotanus et al. [20] have shown that, using this correction term, reliable flux measurements are possible.

The sonic covariance measurements do not produce proper heat flux values if the term $0.51\ \bar{T}\ \overline{w'q'}$ is not negligible compared with $\overline{w'T'}$. However, for characterisation of the atmospheric stability, we are not just interested in the heat flux, but rather in the buoyancy flux, to which the humidity flux also contributes. For this reason the term $\overline{w'T'} + 0.61\ \bar{T}\ \overline{w'q'}$ appears in the definition of the Monin–Obukhov length as given above. The covariance obtained from the sonic anemometer measurement represents $\overline{w'T'} + 0.51\ \bar{T}\ \overline{w'q'}$, showing close resemblance to the term in the definition of L. So if we use directly the sonic anemometer covariance in the calculation of L, the absolute error, $0.1\ \bar{T}\ \overline{w'q'}$, is small. If $\overline{w'T'}$ and $\overline{w'q'}$ have opposite signs, which might occur specifically at the Thorney Island site because of the sea–land transition, the relative error in the buoyancy flux may become large; but this will only occur if the buoyancy flux is very small anyhow, in nearly neutral stability.

The covariance obtained from sonic anemometers may therefore be used directly, without making large errors, for the determination of L. This removes the need for humidity flux measurements, which turns the fact that the heat flux is not directly measured with sonic anemometers into an obvious advantage for this application.

3. What does a dispersion model predict?

Given the input parameters derived from the meteorological data, and other information, a dispersion model can be run to generate predictions of

cloud behaviour. It is then possible to compare with the experimental data. However, it is important that we should know exactly what the model is intended to predict.* Many models would claim to predict average values of concentration. In particular, models which integrate equations of motion over a full-three dimensional grid provide detailed predictions of mean velocity and concentration. But it is difficult to define a suitable average for instantaneous spills. A long time average clearly is not appropriate for instantaneous spills, which are essentially time-varying. A shorter moving average (running mean) smooths the fluctuations in the data; but it also obscures the transient structure of the cloud. Take the example of the concentration measurements shown in Fig. 3, again from Maplin Sands spill 63. The two peaks at the start are associated with the passage of the "roller" at the front of the cloud over the sensor. Such single or double peaks, depending on the elevation of the sensor relative to the height of the roller, are observed at many locations. They cannot be regarded as turbulence; they are part of the *organised structure* of the cloud. This transient behaviour is one of the most interesting and perhaps most important features of the cloud, and should not be smoothed away.

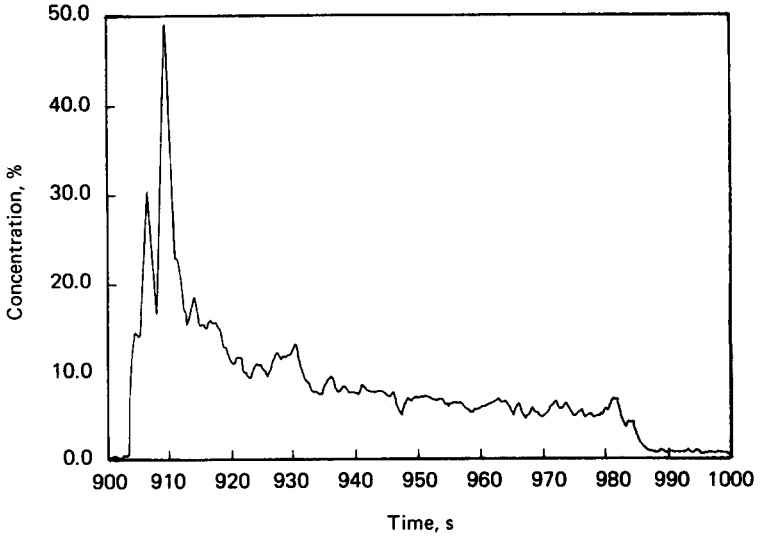

Fig. 3. A gas concentration measurement obtained in Maplin Trial 63 at 64 m from the spill point. The two large peaks at the start are caused by the "roller" at the front of the cloud passing over the sensor.

The only well-defined average which seems to be relevant is an ensemble average, where the ensemble consists of repeated releases with the same wind speed and direction and atmospheric stability. The concept of an "ensemble" of all possible examples, or "realizations", of dispersion under specified con-

*Such a question also arises in comparing steady-state spill data with model predictions. The different considerations in that case have been discussed by Puttock et al. [5].

ditions is discussed by Chatwin [21]. Note, however, that the ensemble considered here is different from the ensemble suggested by Chatwin for use in hazard assessment [21, 22], which includes the statistical variation of wind speed and direction etc. at a particular site. Venkatram [23] has used the idea of an ensemble defined by the input parameters of a model. However, there are difficulties associated with modelling the average over such an ensemble. First, it is not practical to perform a series of field experiments under identical conditions in order rigorously to test the model. To approach this would involve time and expense at least an order of magnitude greater than applied to the Thorney Island trials. But secondly, leaving aside questions of verification, we would suggest that this is not a good basis for a model. The spreading cloud formed by a dense gas release has very sharp changes of concentration and density, not least at the spreading front. In successive releases, even under the same gross atmospheric conditions, the position of the front at a given time will vary in response to eddies in the atmosphere. The result of averaging the result over many releases would be that sharp changes are smoothed out. Even if a model were capable of predicting this smooth average cloud at one stage of its development, there is no reason why a correct physical description of the subsequent development of a smooth cloud would represent the development of the real clouds with their sharp interfaces.

To reinforce this point, let us consider the predictions from the point of view of a model user. (We assume that reliable predictions of probability density functions of concentration in dense gas dispersion, while most desirable, are as yet a distant prospect.) Take the hypothetical example shown in Fig. 4. Curves 1, 2 and 3 represent the concentration at, say, a fix-

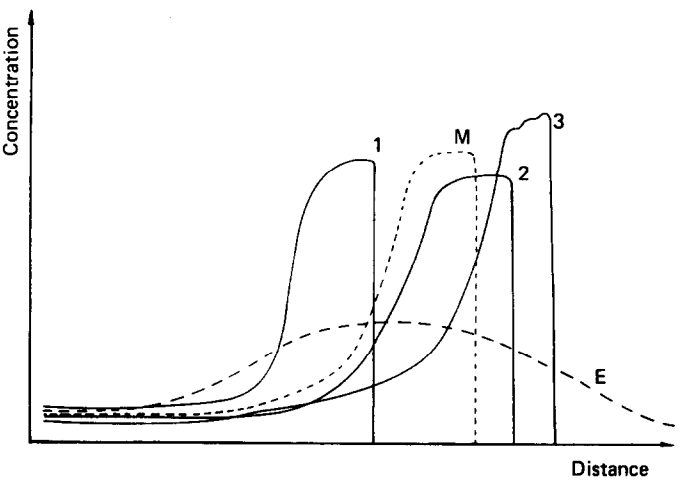

Fig. 4. Hypothetical concentration profiles to illustrate a point about ensemble averaging. 1, 2, 3: measurements in three realisations from an ensemble of releases in the same gross atmospheric conditions. E: the ensemble average over many such realisations. M: a useful model prediction.

ed angle to the wind direction, height above the surface, and time after spill in a series of releases under the same gross atmospheric conditions (realisations from the same ensemble). Curve E represents the ensemble average, obtained from many profiles like 1, 2 and 3. Suppose that a model was able to provide a prediction as shown by curve M. We would suggest this would be more useful than the ensemble average E to a model user trying to assess the consequences of such a release.

It therefore seems that the most useful approach may be to regard a model run as a prediction of a "typical" cloud in some sense which is not rigorously defined. If the data reveal similarities in structure between experiment and model, this provides encouragement that the physics of the dispersion is being modelled correctly.

In addition to studying the details, we can make a number of comparisons of more general observed parameters. As an example Figure 5 shows the maximum concentrations obtained from three different sensor heights as a function of distance from the spill point in Trial 7.

Further examples of such overall characteristics of the cloud have already been presented in this symposium, the time history of area- and volume-

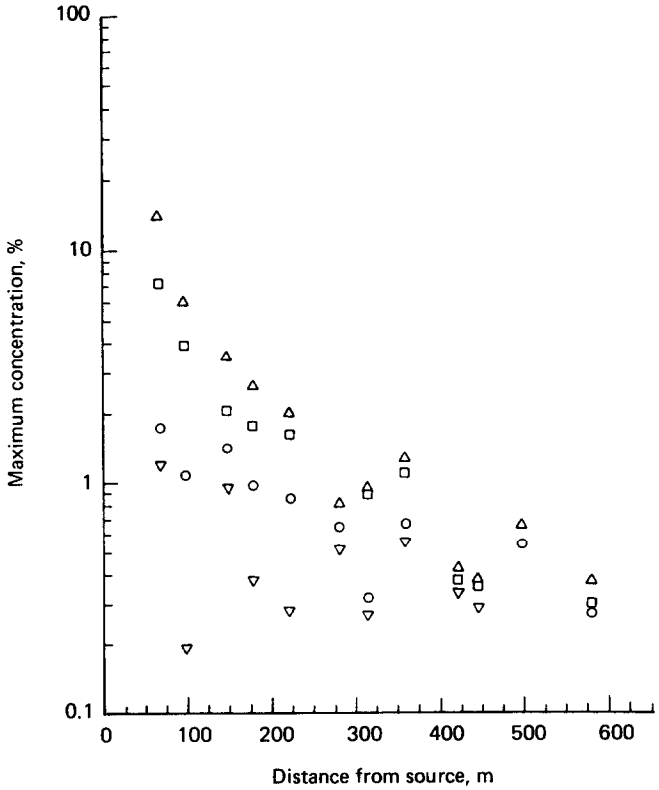

Fig. 5. The maximum gas concentration observed at various locations in Thorney Island Trial 7, plotted against distance from the source.

averaged concentrations, calculated by Brighton et al. [24], provides a view of the development of the cloud. The results of such analyses allow an assessment of the extent to which the entrainment relations in box models represent the area-averaged behaviour of the cloud. A separate look at the data is then necessary to discover how much individual concentration measurements can deviate from the area-average, both as a result of large scale structure and of fluctuations.

Yet another type of analysis [25, 26] allows examination of the bulk motion of the cloud and the spreading due to gravity — further features of the models which can be checked directly.

4. How representative is one realisation?

Another factor to be considered, when assessing the performance of a model in comparison to the data, is the variability of dispersion in the atmosphere between different occasions with the same overall conditions [23]. If only a small number of experimental releases can be performed, then the model has to be compared with one realisation out of an ensemble of possible releases in the same wind speed etc. The model cannot be expected to provide greater precision in its predictions than the inherent variability of the dispersion.

The extent of this variability may be indicated by some results from the Maplin trials. In six steady-state liquid propane releases, the peak concentration observed on each arc of sensors was plotted against distance from the source. From the graph the distance for dilution to 2.1% concentration (the lower flammable limit of propane) was found [5]. The values cannot be compared with each other directly because of differences in both wind speed and spill rate. However, comparisons with predictions from the dispersion model HEGADAS give the results shown in Table 2 for the ratio of observed to predicted distance. The scatter of values could of course be due to deficiencies in the model. But we would suggest that this mostly reflects the inherent variability from occasion to occasion of dispersion, observed over a period of a few minutes, in given mean wind speed and atmospheric stability.

In Phase I at Thorney Island there were two pairs of trials performed on the same day, although in each case unfortunately there were small, but possibly significant, changes in the ambient conditions between the trials. For several other pairs of trials the gross conditions were very similar. So there is a good chance of obtaining an idea of the importance of variability in such releases. At high windspeed such a pair with similar conditions is well provided by Trials 13 and 18. At low windspeed, it will be interesting to see to what extent the observations from Trial 6 lie between those for Trial 7 and Trial 8.

We might, in fact, expect a good agreement between releases at low wind speed, at least in the early stages of dispersion. The flow here is dominated by the slumping of the cloud itself, and the mixing which takes place is largely due to turbulence generated by this flow.

In high winds, however, atmospheric turbulence dominates the dispersion much earlier. Turbulent eddies of all scales up to and exceeding the cloud size are present. Thus both the small scale mixing and the development of the shape of the cloud are more variable.

TABLE 2

For six continuous propane spills at Maplin Sands, the ratio between observed and predicted distances for dilution to 2.1% gas concentration

Trial No.	Spill rate (m^3/min)	Wind speed (m/s)	Distance ratio for LFL, observed/model
43	2.3	5.5	0.88 ± 0.08
46	2.8	8.1	1.02 ± 0.16
47	3.9	5.6	0.73 ± 0.08
49	2.0	6.2	1.28 ± 0.12
50	4.3	7.9	0.79 ± 0.14
54	2.3	3.8	1.53 ± 0.24

5. Conclusions

The Thorney Island Phase I trials have provided a large quantity of data suitable for the validation of mathematical models of dense gas dispersion. In a number of ways, these trials are complementary to other recent dense gas dispersion experiments, which have involved cryogenic liquids.

The meteorological data from such trials, with imperfectly steady and uniform conditions, require careful analysis to provide appropriate model input parameters. Averaging times for the determination of (co-)variances should be taken as long as possible to minimise random errors. Systematic errors in the flux measurements are likely to be significant, but correction for them is quite straightforward.

A well-proven method to characterise turbulence and stability in the atmospheric surface layer is determination of the surface layer parameters defined in the Monin—Obukhov similarity theory. The sea—land transition and measurement errors make it necessary to use at least two methods to determine these parameters from the measurements, thus allowing for a check on consistency. Two such methods are profile fitting and direct flux measurements.

In comparisons of data with predictions of a dispersion model, a clear idea of what the model is intended to predict is needed. Time averaging or even ensemble averaging may obscure important features of the data. The most useful three-dimensional model predictions may be those which correspond well with typical observations, rather than those which coincide with a simply-derived average of data.

It is also important to take into account the variability of dispersion in

the atmosphere between occasions with the same gross atmospheric conditions. However, this may be less of a problem in strongly density-driven flows than for neutral releases.

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