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Methods of deriving input met parameters for ADMS from the types of simple measurements available at nuclear sites.

by

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Notation

To avoid interrupting the flow of the text with definitions, we list here the main notations used. We have where possible tried to follow the ADMS notation conventions.

- A: solar declination (in degrees)
- c_p: specific heat capacity of air ($\simeq 1004.6$ J kg^{-1} K^{-1})
- c_l: cloud cover (in oktas)
- c_{lm}: 'modified' cloud cover (in oktas)
- $F_{\theta 0}$: atmospheric flux of sensible heat at the ground
- g: acceleration due to gravity ($\simeq 9.81$ m s^{-2})
- h: boundary layer depth
- k: von Karman's constant ($\simeq 0.4$)
- K^+: incoming solar radiation
- L_{MO}: Monin-Obukhov length
- N_u: buoyancy frequency above the boundary layer top ($= \sqrt{g\gamma_\theta/T_0}$)
- P: precipitation rate

- $\mathcal{P}$: Pasquill stability category
- Q_*: net radiation
- r: surface albedo
- s: sine of the solar elevation
- S_{lat}: latitude (in degrees)
- t_{day}: Julian day number
- t_{hour}: local time in hours
- T_0: absolute temperature
- U: wind speed
- u_*: friction velocity
- z_0: roughness length
- α: modified Priestley-Taylor parameter, characterising the amount of water available for evaporation.
- γ_θ: rate of increase of potential temperature with height above the boundary layer.
- ΔT: near surface temperature over land minus sea surface temperature
- $\Delta\theta$: temperature jump across the boundary layer top
- ϕ: wind direction
- ρ_a: air density ($\simeq 1.225$ kg m^{-3})
- σ_w: standard deviation of the vertical velocity
- $\sigma_{y/z}$: plume spread in the crosswind and vertical directions
- σ_θ: wind direction standard deviation
- θ_*: surface layer temperature scale
- $\psi(\zeta)$: Monin-Obukhov similarity function

1 Introduction

A number of models exist for calculating the dispersion of material in the turbulent layer near the ground (the 'atmospheric boundary layer' – typically between 50 m and 2 km in thickness) over short ranges downwind from a source. Short range here means distances over which changes in meteorology can be neglected in considering the movement of any one air parcel over the distance. Typically such models will be useful up to 30 km downwind. The Atmospheric Dispersion Modelling System (ADMS) is a new model of this type (Carruthers et al 1992, 1994) which treats met data in a rather different way to older models such as R91 (Clark, 1979). In particular it abandons the idea of using Pasquill stability (Pasquill 1961). In this report we investigate methods of deriving input met parameters for ADMS from the types of simple measurements available at nuclear sites. Throughout this report, ADMS refers to ADMS version 2.02 (dated 8/12/95).

2 Overview of met requirements for short range models such as R91 and ADMS

In order to predict the dispersion of material over short ranges in the atmospheric boundary layer some information is required on the air flow and turbulence. A wind speed and direction measurement from a sufficiently well exposed anemometer is generally sufficient to determine the air flow, at least over relatively flat homogeneous terrain. This is because it is possible to make theoretical estimates of changes with height to sufficient accuracy for practical purposes. In many situations the changes are in any case not great and some models do not even attempt to estimate such changes and assume that the wind direction and/or wind speed is uniform with height. R91 assumes constant wind speed and direction while ADMS assumes constant wind direction but models the variation in wind speed with height.

Obtaining information on the turbulence is more difficult. Turbulence is rarely measured directly except in research experiments and certainly not at a range of heights spanning the depth of the boundary layer. Also the variation of turbulence with height can be important and so measurements at a single height have a limited value. As a result it is necessary to estimate properties of the turbulence from other quantities which are known. Modern models, such as ADMS, estimate turbulence levels (variances of the fluctuating velocity components) and turbulence time-scales from known quantities and then use the levels and time scales to estimate the turbulent spread of a plume. Older models, such as R91, estimate the spread of a plume due to turbulence via an intermediate quantity called Pasquill stability, rather than estimating turbulence levels as an intermediate step. In all cases however we need to estimate something about the turbulence or its dispersive characteristics from what is known, and to do this we need to understand the sources of turbulence.

In the atmospheric boundary layer, turbulence arises from two sources, wind shear and surface heating by the sun. In the first case the turbulence draws its energy from the mean flow ('mechanically generated' turbulence) while in the second the turbulence arises from convective overturning of the air which is driven by the warming of the ground and of the air in contact with it. At night the ground cools by radiation to space and so the presence of cool air near the ground acts to suppress the mechanically generated turbulence. For dispersion purposes one needs a way of characterizing the relative importance of thermal turbulence generation (which can be positive or negative) and of mechanical generation (which is always positive). The reason that it is the *relative* importance of these two sources of turbulence that one needs to estimate is that if one doubles the wind speed (and so the velocities of mechanically generated eddies) and also doubles the velocities of the convectively generated eddies then the plume will spread twice as fast in time, but travel downwind twice as quickly, and so the plume width at a given downwind distance will be unaltered. It is also worth considering what happens if the wind speed changes without altering the convective turbulence generation. If conditions are very convective, then the turbulence will be dominated by the convectively generated turbulence and so will not alter as the wind speed changes. As a result, the rate of plume spread in time will not change if the wind speed is doubled, but the plume will reach a given downwind distance twice as quickly and so, at the given downwind distance, will have a smaller spread. Of course, if the wind speed continues to increase, the mechanical turbulence will eventually become important and the plume spread (at the given downwind distance) will not continue decreasing.

Older models such as R91 use the 'Pasquill stability' $\mathcal{P}$ to characterize the relative importance of mechanical and thermal effects. In Pasquill's original paper (Pasquill 1961) Pasquill stability is calculated from wind speed U and insolation K^{+} (incoming solar radiation) during the day and wind speed and cloud cover at night, cloud cover being a surrogate for heat loss from the ground to space. The method used is as follows:

Table 2.1	Daytime			Nighttime	
	Insolation			Thinly overcast or	
U (ms^{-1})	Strong	Moderate	Weak	low cloud ≥ 4 oktas	cloud ≤ 3 oktas
< 2	A	A-B	B	-	-
2-3	A-B	B	C	E	F
3-5	B	B-C	C	D	E
5-6	C	C-D	D	D	D
> 6	C	D	D	D	D

In applying the method during the day, the insolation has to be estimated and this is usually done using cloud cover, time of day and time of year. A problem with Pasquill stability is that it doesn't enable one to estimate changes in turbulence levels with height. Also a number of alternative schemes to Pasquill's original scheme have been devised and, because one cannot measure Pasquill stability, there is no objective way to say which is better (except by comparing dispersion estimates from Pasquill stability based dispersion models with experimental dispersion data – but this is likely to show that different definitions of Pasquill stability work best with different models). The differences between such schemes are not small (see Appendix A below).

Because of the above problems, and also because the current understanding of turbulence and dispersion in the boundary layer doesn't involve Pasquill stability, modern models such as ADMS tend to characterize the relative importance of thermal and mechanical turbulence generation through the Monin-Obukhov length, L_{MO}, defined by

$$L_{MO} = -\frac{\rho_a c_p T_0 u_*^3}{k g F_{\theta 0}} \tag{1}$$

where ρ_a is the air density ($\simeq 1.225$ kg m^{-3}), c_p the specific heat capacity of air ($\simeq$ 1004.6 J kg^{-1} K^{-1}), T_0 the absolute temperature, k von Karman's constant ($\simeq 0.4$), g the acceleration due to gravity ($\simeq 9.81$ m s^{-2}), $F_{\theta 0}$ the atmospheric flux of sensible heat at the ground and u_* is the friction velocity, characterising the drag of the ground on the atmosphere. Since it is the absolute temperature, T_0 does not need to be known very accurately – taking $T_0 \simeq 288$ K, i.e. 15°C, is probably adequate. Also if u_* is not measured (as is usually the case) it can be estimated from the wind U at some height z, $F_{\theta 0}$, and the surface roughness length z_0 (which is a function of the terrain type). Finally, if $F_{\theta 0}$ is not measured (again this is usually the case), it can be estimated using a variety of quantities. A number of schemes exist for estimating $F_{\theta 0}$ (see e.g. Galinski and Thomson (1995)). In many of these schemes, the quantities used for estimating $F_{\theta 0}$ include cloud cover, time of day and time of year (compare estimation of insolation in calculating Pasquill stability – $F_{\theta 0}$ is not of course quite the same as insolation K^+, the difference being due to reflection of solar radiation from the ground, long wave radiation from the ground and atmosphere, ground heat flux, and the energy implications of evaporation or dew formation). As is implied above, direct measurements of u_* and $F_{\theta 0}$ are unusual but are possible using 'eddy-correlation' techniques (see e.g. Kaimal and Finnigan (1994)). This requires a means of measuring the turbulent fluctuations in wind (including vertical velocity) and temperature. L_{MO} in fact characterizes turbulence in a way which is not that dissimilar from $\mathcal{P}$ in that, for given z_0, there is a rough correspondence between $\mathcal{P}$ and L_{MO} (Golder 1972). However the correspondence is not exact. Unlike $\mathcal{P}$, L_{MO} is dimensional and $|L_{MO}|$ indicates the height at which thermal turbulence generation becomes important. (Note this means that, although L_{MO} tells us the relative importance of mechanical and thermal turbulence generation *at a given height*, to estimate the relative importance in the boundary layer *as a whole* one also needs to know the boundary layer depth h. This is the reason h/L_{MO} is often regarded as an important stability parameter.) By convention L_{MO} has the opposite sign to $F_{\theta 0}$, being negative in unstable conditions and positive in stable conditions. Near the ground mechanical turbulence generation is always dominant. The variation of L_{MO} with stability is slightly strange with L_{MO} being small and positive in strongly stable conditions, small and negative in strongly unstable conditions, and large in magnitude (positive or negative) in near neutral conditions (see figure 2.1). $1/L_{MO}$ is better behaved and varies continuously with stability.

In addition to an estimate of the relative importance of mechanical and convective turbulence generation, an estimate of the boundary layer depth h is usually required. In most models (including R91 and ADMS) this provides an upper limit to mixing. In ADMS it also fulfils a further role since it is used in determining turbulence levels. The main effect of h on the turbulence is to influence the variation of turbulence properties with height with, in general, near surface properties being independent of h. However, in convective conditions, large eddies of scale comparable to the boundary layer depth provide a significant contribution to horizontal turbulent velocities near the ground, and so horizontal turbulence depends on h, even near the ground. In certain cases dispersion predictions can be very sensitive to h, e.g. if changing h means that the source height (or the effective source height due to buoyant plume rise) moves from being above to being within the boundary layer, but more generally the sensitivity, although significant, is not expected to be very great. For example, in its role as an upper limit to mixing, h has little effect close to the source for sources well within the boundary layer and results in concentrations proportional to $1/h$ at longer range when the pollutant is well mixed in the vertical. Also its effect on the variation of turbulence with height is expected to be very significant only for sources in the upper half of the boundary layer. Finally, in its role of determining the characteristics of horizontal eddies, typical horizontal turbulent velocities depend on h only as $h^{1/3}$. The dependence of h on other boundary layer properties is quite complex in general with h depending not just on local properties but on the history of the air mass. In particular, during convective conditions h tends to grow in time at a rate depending on the stability above the boundary layer and the sources of turbulence U and $F_{\theta 0}$.

Over non-homogeneous terrain more extensive data would be desirable in principle. However the simplicity of models like R91 and ADMS means that they cannot make use of very extensive data. Perhaps the most important thing which could be measured is the wind direction at the release height. As an example, consider a situation with a sea breeze at a coastal site. In such situations the direction of the wind aloft could be very different from that at the surface. Somewhat special consideration has to be given to input winds in ADMS if the hills, coastline or building effects modules are used, taking account of the way the met data is treated by ADMS (e.g. does ADMS expect as input a wind representing what the wind would be in the absence of the hill, coast or building?). This is discussed in detail in §3 below. However, even in this case, ADMS can only accept a single wind speed and direction value as input. To do more is difficult without a model which can resolve spatial and temporal changes in meteorology and calculate their evolution, for example the UK nuclear accident response model which is based on the Met Office's numerical weather prediction models (Ryall et al, 1995).

3 Details of ADMS's met requirements

In this section we set out in detail the input variables which ADMS requires or would like. ADMS can consider several sets of met data in a single run and will calculate dispersion separately for each. These sets of data might consist of sequential hours of data or constitute some more general collection of cases.

(i) Roughness length, z_0. This is not really a met parameter but is used extensively in processing the met data and so is mentioned here for completeness. A value of z_0 is essential for ADMS to run.

(ii) A wind speed U and direction ϕ at some specified height z or the geostrophic wind speed and direction (in principle the friction velocity u_* could be given as an alternative to $U(z)$, but this option is unlikely to be used much in practice). Input of these variables is essential. If the coastline module is used, ADMS assumes that the value is a 'far-inland' value. However ADMS only includes thermal internal boundary layer effects in the coastline module. Sea breeze effects are not included and so the only way to include such effects is to alter the input wind to account for them. If the hills or buildings effects modules of ADMS are used, then ADMS assumes the input wind is an 'upstream' value, unaffected by the hill or building.

(iii) Monin-Obukhov length L_{MO}. If L_{MO} is input to ADMS, it is used directly. If L_{MO} isn't (which will usually be the case) but $F_{\theta 0}$ is, ADMS will calculate L_{MO} from $F_{\theta 0}$, T_0, $U(z)$ (or u_* if this is input to ADMS – see (ii) above) and z_0, with T_0 approximated by 288.15K (15°C) if T_0 is not input. Finally, if neither L_{MO} nor $F_{\theta 0}$ is input but c_l, t_{hour} and t_{day} are, ADMS will calculate $F_{\theta 0}$ (and thence L_{MO}) from z_0, $U(z)$ (or u_*), c_l, T_0, r, α, t_{hour} and t_{day} using defaults for T_0, r and α if these are not input. The calculation method is explained in Appendix B. It is essential that ADMS can estimate L_{MO} and so it is essential that either L_{MO} or $F_{\theta 0}$ or c_l, t_{hour} and t_{day} are input. ADMS always calculates u_* and $F_{\theta 0}$ even if this is not necessary to obtain L_{MO}; this is because they may be used elsewhere (e.g. in calculating boundary layer depth as discussed below). ADMS cannot use Pasquill stability $\mathcal{P}$ directly to estimate L_{MO}. However, a possible way of using such data is discussed in Appendix A.

If the essential requirements of ADMS described in (i), (ii) and (iii) are not met, the dispersion is not calculated and ADMS *should* then move on to the next set of met data, although sometimes ADMS 2.02 fails in such cases.

(iv) Boundary layer depth h. If h is not specified, ADMS can estimate it. When $F_{\theta 0} \leq 0$ (stable conditions), ADMS estimates h using only u_* and L_{MO}. When $F_{\theta 0} > 0$ (unstable conditions) however, ADMS can produce a scientifically sounder estimate if some extra data is available. As noted in §2, h grows during the day at a rate dependent on the lapse rate above the boundary layer and the sources of turbulence U and $F_{\theta 0}$. The lapse rate above the boundary layer is not something that can be input directly into ADMS; however the buoyancy frequency above the boundary layer N_u can be input and ADMS will calculate the lapse rate from this or, more usually, from a default value of N_u. To calculate h, ADMS tries to estimate, or adopt defaults for, the history of u_*, $F_{\theta 0}$, N_u and T_0 back to the last stable occasion ($F_{\theta 0} < 0$) in order to apply a model for the evolution of h. If it fails to estimate the history, a neutral value of h is used. The procedure for estimating h is illustrated in figure 3.1 The method used to estimate the history of u_*, $F_{\theta 0}$, N_u and T_0 takes two forms, depending on whether or not the data input into ADMS consists of a series of sequential hours of data. If it does consist of sequential hours, ADMS can make use of the data input for previous hours as well as for the current hour. For sequential data the method used is as follows:

- If N_u and/or T_0 not input for previous hours, set to default value
- If u_* unavailable for previous hours (i.e. not input and not estimatable as in (iii) above), interpolate in time if possible
- If u_* still unestimated for previous hours, set to first available subsequent value (up to and including the current hour)
- If $F_{\theta 0}$ unavailable for previous hours (i.e. not input and not estimatable as in (iii) above), interpolate in time over any gaps no longer than two hours
- If $F_{\theta 0}$ still unestimated for previous hours, but c_l, t_{hour} and t_{day} available for the current hour, estimate $F_{\theta 0}$ on the basis that c_l, r and α are unchanged in the previous hours.

For non-sequential data the following approach is adopted:

- Set u_*, N_u and T_0 to the values in the current hour
- If c_l, t_{hour} and t_{day} available for the current hour, estimate $F_{\theta 0}$ for previous hours on the basis that c_l, r and α are unchanged in these hours.

(v) Properties above the boundary layer, N_u and $\Delta\theta$. In ADMS the state of the atmosphere at and above the boundary layer top is characterised by N_u and $\Delta\theta$. As noted above N_u also has a role in determining h. These values will usually be determined within ADMS but can be specified by the user.

(vi) Wind direction standard deviation, σ_θ. ADMS can make use of information on the wind direction standard deviation in calculating lateral spread. However, the value required is that due to changes in 'meteorological' wind direction, with ADMS adding in the spread due to turbulence. If not input, ADMS will estimate the size of changes in meteorological wind direction from the 'sampling time' for the concentration measurement (Thomson 1995).

(vii) Difference between near surface temperature over land and sea surface temperature, ΔT. This is required by ADMS if the coastal module is to be used. The two temperatures can also be input separately – ADMS will then calculate the difference.

(viii) Precipitation rate, P. This is generally needed if ADMS is to calculate wet deposition (ADMS can calculate wet deposition without precipitation information by assuming a wash out rate independent of rainfall rate. Clearly this only makes sense in a statistical sense – without precipitation information it makes little sense to estimate wet deposition on any individual occasion).

4 Instrumentation available at nuclear sites

During 1993 the Met Office arranged inspection visits to all nuclear sites in the UK to assess the quality and representativeness of meteorological observations made at the sites in support of PACRAM (Procedures And Communications in the event of a release of RadioActive Materials – Met Office (1995)).

All sites have a low level anemometer at a height in the region of 10 to 20m and just over half the sites also have a high level anemometer at a height which is between 30 and 75m with the exception of Sellafield where it is at 120m. Most sites have some means by which the variability of wind direction could be estimated. Additional data vary between sites. Pressure, rainfall, various temperatures (e.g. dry bulb, wet bulb and ground temperatures) and humidity are the most common measurements. Also sunshine is measured at Torness and Sellafield and there is a solarimeter at Torness. In an emergency cloud amount would be available from most sites but the accuracy of such estimates may be uncertain if, for example, trained observers are not available or if there is limited vision from within the control room. Cloud observations are made routinely at Dungeness, but it is not clear whether this is so at other sites.

5 Meeting the ADMS data requirements from the type of simple measurements available at nuclear sites

Here we consider the needs and desires of ADMS for met data as outlined in §3 and how these can be satisfied by the types of data available as summarised in §4. In fact the limited range of data available means that the options are not very numerous. We will consider ADMS's needs in the eight categories of data described in §3. A number of ADMS calculations have been carried out to investigate model sensitivity to met inputs. In all these cases we assumed a latitude of 52°N, a roughness length of 0.1m, an averaging (or 'sampling') time of 1 hour, and a continuous passive release with source diameter of 1m. We also used default values of met variables and a release height of 30m except where explicitly mentioned. It has been impossible within the scope of this project to conduct sensitivity tests across the whole range of met and source conditions. The results must be interpreted with this in mind; although indicative, they cannot be used to make firm conclusions about the importance of some parameter under *all* conditions.

(i) Roughness length z_0. Strictly speaking z_0 is not a met parameter and so its estimation lies outside the scope of this report. However, because of its use in processing the met data we include some brief comments. z_0 can be estimated from some types of met measurements (e.g. turbulence measurements or profiles, although profiles can be difficult to interpret except over very homogeneous terrain) but these lie outside the scope of the simple measurements available at nuclear sites. As a result z_0 must be estimated from a general qualitative description of the terrain. For consistency we recommend following the values recommended in the ADMS help screen:

- Cities, Woodland 1.0m
- Parkland, Open Suburbia 0.5m
- Agricultural areas 0.2 to 0.3m
- Root crops 0.1m
- Open grassland 0.02m
- Short grass 0.005m
- Sandy desert 0.001m

However, it should be noted that it is rare in this country to get extensive areas with roughness lengths less than 0.1m.

(ii) Wind speed U and direction ϕ. Because changes in wind speed with height are not always in accord with the theory used in ADMS (e.g. due to sea breezes, baroclinicity, topography, random variability, or indeed errors in the theory) and because ADMS does not account for changes in direction with height (except changes due to terrain when the complex terrain effects module is used) it is generally best to always use a wind speed and direction measured as close to the release height as possible (or possibly higher for buoyant releases) subject to being far enough above the surface roughness elements for the anemometer to be properly exposed. There are, however, two reasons why one might not wish to do this.

The first reason concerns the fact that, as noted in Appendix B, using a high level input wind imposes a limit on how stable ADMS can estimate the flow to be. To investigate this we ran ADMS for clear sky night-time conditions with wind speeds of 1, 2, 3, 4 and $5ms^{-1}$ at 10m. From the output of ADMS's met input module (which includes values of u_* and L_{MO}) we calculated what ADMS estimated the wind speeds to be at 50m (about the height of the upper wind measurements at most nuclear sites) and reran ADMS using these 50m winds. The results are as follows:

Table 5.1

U at 10m (ms^{-1})	Values from 10m run u_* (ms^{-1})	Values from 10m run $1/L_{MO}$ (m^{-1})	U at 50m computed from 10m run (ms^{-1})	Values from 50m run u_* (ms^{-1})	Values from 50m run $1/L_{MO}$ (m^{-1})
1	0.0149	1.7741	2.915	0.0381	0.3973
2	0.0400	0.7646	4.335	0.0625	0.3142
3	0.1924	0.0331	6.144	0.1925	0.0331
4	0.3021	0.0134	7.040	0.3017	0.0135
5	0.3993	0.0077	8.074	0.3996	0.0077

For the three highest wind speeds the results are virtually identical. However, for the two lower wind speeds, the ADMS runs with the 50m wind pick up a different solution for u_* and L_{MO} (as discussed in Appendix B) which is more neutral than that obtained from the 10m wind. This suggests that it might be preferable to use the lower level wind even for higher level releases. However, these cases are associated with quite stable flows where ADMS predicts that the plume grows very slowly in the vertical. For these cases ADMS predicts that passive releases at a height of about 50m (which is close to the boundary layer top for the most stable cases) do not reach the ground in significant amounts by 30km downwind and releases at a height of 30m do not reach their maximum ground level concentration before 30km. In such cases there is a lot of uncertainty in the rate of spread and so the results have a high degree of uncertainty. As a result we judge that using a wind measurement near the release height is still preferable.

The second reason concerns possible use of the hills module in ADMS. As noted in §3, this module assumes that the input wind is a wind undisturbed by the topography, and so it might be preferable to use a wind measurement some distance away from the source, even if this is not at the release height. Unfortunately it is hard to make general recommendations here, as the best procedure is likely to be somewhat case specific. In any case the measurements made at nuclear sites do not generally involve more than one location and so no choice is possible (unless measurements not associated with the site are used). The best solution here would be for ADMS to be modified so that it can make effective use of topographically disturbed wind measurements.

(iii) Monin-Obukhov length, L_{MO}. For the majority of the nuclear sites the currently available data is such that there is no option but to use cloud cover, time of year and time of day to calculate L_{MO}. Because it isn't generally clear whether cloud is measured routinely (i.e. other than in emergency) or reliably we have investigated the utility of using cloud observations from nearby met sites as a substitute for on site measurements. Hollis (1995) analysed cloud cover observations at two nearby sites, Finningley and Waddington, which are separated by about 40km. He found that on 77% of occasions the cloud cover reported at the two sites differed by 1 okta or less. It seems unlikely that such errors would be very significant for ADMS, although it is hard to test this across the whole range of release heights and met conditions. To partially investigate this, we performed some ADMS runs for mid summer conditions with varying cloud amount and with $t_{day} = 182$ and $U(10\text{m}) = 3\text{ms}^{-1}$. The values of $1/L_{MO}$ calculated by ADMS are as follows:

Table 5.2

c_l	$1/L_{MO}$ (m^{-1}) at $t_{hour} = 12$	$1/L_{MO}$ (m^{-1}) at nighttime ($t_{hour} = 0$)
0	-0.0529	0.0331
1	-0.0534	0.0326
2	-0.0537	0.0312
3	-0.0533	0.0289
4	-0.0516	0.0260
5	-0.0479	0.0226
6	-0.0412	0.0190
7	-0.0300	0.0152
8	-0.0107	0.0115

These values show that the consequent changes in $1/L_{MO}$ of an error in cloud amount of 1 okta are not great, except for daytime conditions with high values of cloud amount. Even here the consequent changes in ADMS's σ_y and σ_z are not great (see fig 5.1) although they are not negligible in the case of σ_z. The use of cloud from a nearby site seems a viable option, although on occasion it will result in small (but not negligible) errors. If cloud from a nearby site is used, care must be taken to ensure the climatology is reasonably similar; for example cloud at coastal and inland sites could show larger differences from those found at Finningley and Waddington.

We have also investigated the extent to which information on T_0, r and α (which can be used in calculating $F_{\theta 0}$ and L_{MO} from c_l, t_{day} and t_{hour} if data is available) is useful. If we ignore the fact that T_0 has an effect on the relation between L_{MO}, $F_{\theta 0}$ and u_* through equation (1) (as noted above, this effect is small because T_0 is the absolute temperature), then T_0, r and α only have an effect during the day. To test the effect of T_0 we ran four cases, two mid-winter and two mid-summer, with varying T_0. Other met variables were $U(10\text{m}) = 3\text{ms}^{-1}$, $t_{hour} = 12$ and $c_l = 0$. The values of $1/L_{MO}$ calculated by ADMS are as follows:

Table 5.3

t_{day}	temperature (°C)	$1/L_{MO}$ (m^{-1})
1	-5	-0.0113
1	5	-0.0074
1	15	-0.0037
182	15	-0.0529
182	25	-0.0375

The effect of T_0 on $1/L_{MO}$ is seen to be significant, although not dramatic, but the consequent effect on σ_y and σ_z is of significance only for σ_z in the mid-summer case (see figure 5.2). We conclude that information on T_0 is useful but not essential.

The main cause of changes in r is increased reflection from lying snow. Hence we have performed two mid-winter runs with r corresponding to the ADMS recommended value for snow covered ground and to the ADMS default value. Other met variables were $U(10\text{m}) = 3\text{ms}^{-1}$, $t_{hour} = 12$, $c_l = 0$ and $t_{day} = 1$. Values of $1/L_{MO}$ from ADMS are as follows:

Table 5.4

r	$1/L_{MO}$ (m^{-1})
0.23	-0.0037
0.6	0.0235

The effect of changing r is very significant with the stability changing from unstable to stable, and the consequent change in σ_z is very great although σ_y is almost unchanged (see figure 5.3). It is therefore important to assess whether there is substantial snow cover, but it is probably impractical to do more than this, e.g. by allowing r to vary continuously between 0.23 and 0.6.

We also conducted some tests on the effect of changing α for mid-summer and mid-winter with $U(10\text{m}) = 3\text{ms}^{-1}$, $t_{hour} = 12$ and $c_l = 0$. We performed runs with $\alpha = 1.0$ (the ADMS default) appropriate for moist grassland and with a smaller value of 0.5 appropriate to drier conditions. Values of $1/L_{MO}$ from ADMS are as follows:

Table 5.5

t_{day}	α	$1/L_{MO}$ (m^{-1})
1	1.0	-0.0037
1	0.5	-0.0180
182	1.0	-0.0529
182	0.5	-0.0884

Changing α has a significant effect on $1/L_{MO}$ and, for the mid-summer cases, this feeds through to a significant change in σ_y and σ_z (see figure 5.4). This suggests that the use of the default for all conditions is unsatisfactory and a smaller value should be used in drought conditions or urban areas. We recommend $\alpha = 0.5$, although there is not much evidence on which to make a choice of α. Holtslag and van Ulden (1983) found $\alpha = 0.45$ appropriate for the dry grassland of the Project Prairie Grass experiment.

A few sites have solarimeters to measure K^+ or sunshine recorders. These could be used as a substitute for cloud during the day and we have investigated the possibility of doing this. An advantage of such approaches is that measurements of K^+ or sunshine are more easily (and cheaply) automated than cloud measurements. If K^+ is measured, then one could attempt to calculate $F_{\theta 0}$ using the ADMS scheme (appendix B) but by-passing the calculation of K^+. (Although the ADMS *scheme* could be used here, the ADMS *code* cannot be used directly since it is not currently possible to bypass the calculation of K^+. This would be possible in future versions of ADMS if ADMS was changed to allow the user to input K^+ directly.) Cloud amount is still required in estimating Q_* from K^+, but the dependence here is rather weaker than in the calculation of K^+. Using data measured at Cardington, Galinski and Thomson (1995) investigated calculating Q_*, and hence $F_{\theta 0}$, from K^+ by assuming a constant value of cloud amount. Results for $F_{\theta 0}$ were best when $c_l = 13$ was assumed! This is probably because the mean and modal values of cloud cover are quite large (the mode is 7 for Finningley – Hollis (1995)) and that, at least at Cardington, the scheme showed a significant mean bias which was corrected by taking c_l outside its natural range. It is probably dangerous however to draw too much from behaviour at one site and taking $c_l = 7$ seems more sensible. With this value, Galinski and Thomson (1995) found errors in $F_{\theta 0}$ increased by 20% relative to those occurring when cloud is used instead of solarimeter data. This will feed through to about a 20% increase in the error in $1/L_{MO}$. The simulations presented above suggest that this sort of additional error is acceptable.

Table 5.6 shows frequency of cloud amount and sunshine at Finningley (from Hollis (1995)).

Table 5.6

c_l	sunshine duration (tenths of an hour) 0	1	2	3	4	5	6	7	8	9	10
0	21	2	5	8	8	5	2	8	14	15	451
1	43	15	25	22	41	25	33	56	62	87	1411
2	34	26	25	27	43	51	55	76	106	149	810
3	72	53	37	52	62	59	75	90	112	219	655
4	79	42	50	54	58	65	77	75	99	156	390
5	175	93	101	109	99	93	108	140	134	207	422
6	504	232	188	201	153	158	174	176	199	255	489
7	4461	921	616	463	355	311	308	286	269	281	389
8	5755	145	81	35	33	21	15	4	8	5	5

This shows that, by taking $c_l = 1$ when sunshine duration is ten tenths, $c_l = 7$ when sunshine is zero or one tenth, and $c_l = 6$ at other times would give the cloud amount correct to within 1 okta on 78% of occasions, similar to that when cloud from a nearby site is used. When discussing cloud from a nearby site, we judged such an error to be acceptable. However it is not clear that this is so here – the approach suggested will not represent the most extreme conditions (i.e. $c_l = 0$ or 8) and the performance could be much worse in locations with different cloud climatology. This would require more investigation.

Unfortunately solarimeters and sunshine recorders are of no use at night. As a result we discuss the somewhat radical option of diagnosing L_{MO} from $U(z)$ and z_0 alone. Although it initially seems unlikely, such an approach might be possible because of the fact that $F_{\theta 0}$ at night depends much more strongly on wind speed than on cloud cover (see e.g. Tubbs (1988)). Galinski and Thomson (1995) found using Cardington data that assuming a constant value of cloud cover gave comparable accuracy in estimating $F_{\theta 0}$ to that obtained using actual cloud measurements. As above, taking $c_l = 7$ seems appropriate. The fact that this is a viable option is perhaps more to do with the relatively poor performance of schemes for diagnosing $F_{\theta 0}$ from wind and cloud at night, rather than the good performance of a scheme which ignores cloud data! Although the statistical evidence from Cardington supports the use of $c_l = 7$, this seems a little unsatisfactory without an understanding of the underlying physics. Also the approach will not represent the most extreme conditions which seems undesirable. Perhaps a better approach is to use the cloud observations from a nearby met site. Because of the lack of sensitivity to cloud found by Tubbs (1988) and Galinski and Thomson (1995), this is likely to be more appropriate at night than during the day.

A further possibility which may be worth considering is the use of net radiation measurements. Such measurements tell us something about what is happening at night as well as during the day. However, net radiation instruments require more looking after than solarimeters and it is not clear how such information could be used within ADMS at night – some research would be needed to investigate the possibilities here. During the day Q_* data could be used easily by by-passing part of ADMS's $F_{\theta 0}$ calculation in the same way as discussed above for K^+ data. The study by Galinski and Thomson (1995) shows that Q_* is reasonably well predicted from c_l and/or K^+ (most of the uncertainty in predicting $F_{\theta 0}$ is due to partitioning the net radiation between $F_{\theta 0}$, evaporation and ground heat flux) and so there is no particular advantage in having Q_* measurements during the day.

Although direct measurements of heat flux or inferring heat flux from profile measurements of wind and temperature lie outside the 'type of simple measurements made at nuclear sites', it is worth making some brief comments on these. Direct measurements of heat flux can be made using the eddy-correlation method with e.g. a sonic anemometer. Such instrumentation is fairly robust and maintenance free, but measuring $F_{\theta 0}$ this way has more exacting exposure requirements than a simple wind speed and direction measurement and requires either careful instrument levelling to avoid apparent mean vertical velocities or appropriate correction of the data. Other corrections may also be necessary (see e.g. Schotanus et al (1983), Grant and Watkins (1989)). With a sonic anemometer it would be possible to measure u_* as well as $F_{\theta 0}$ and so measure L_{MO} directly. A problem with this is that results can be contaminated by statistical noise in the measurements. During the day it is probably more robust to use U rather than attempting to measure u_* while at night it may be better to measure the standard deviation of the vertical velocity, σ_w, and use this to infer u_* via the relation $\sigma_w \simeq 1.3u_*$ (see e.g. Pasquill and Smith (1983)). Using profile measurements to infer heat flux is generally difficult because of the need to detect temperature and/or wind *differences* and the possibility that results may be contaminated by the non-uniform nature of the site (higher instruments will be affected by surface properties over a wider area). Grant (1996) considered the problem of using temperature at two heights and wind speed at one height over a uniform surface to infer stability. He deduced that temperatures need to be measured to an accuracy of order 0.07°C which would be difficult to achieve routinely. The approach therefore seems impractical for routine use and would in any case require further research, particularly on the issue of contamination by non-uniformities in surface properties.

(iv) Boundary layer depth h. Measuring h is difficult and expensive, requiring measurements aloft or remote sensing. These methods fall outside the type of simple measurements made at nuclear sites. Hence ADMS will have to estimate h from the available data. We assume here that $F_{\theta 0}$ is derived by ADMS from c_l, t_{day} and t_{hour} and consider the question of whether ADMS can benefit significantly during the day from knowing the meteorological history. Figure 5.5 shows the evolution of h as calculated by ADMS for six sample days. The days chosen are 1/1/85, 1/3/85, 1/5/85, 1/7/85, 1/9/85 and 1/11/85 at Wattisham and the input data consists of hourly values of U(10m), ϕ, c_l, T_0, t_{day} and t_{hour}. Values of h based on met data for a single hour and also values which take account of the history of the meteorology are shown (the latter are obtained by telling ADMS that the met data are 'hourly sequential'). A seventh day is also shown which consists of the data for 1/7/85, but with cloud amount set to 8 oktas up to $t_{hour} = 16$ and set to zero at later times. The six sample days show that, although the values of h differ, the differences are not great and so the use of a single hour of data on its own will normally give acceptable results. However on occasions results can be bad. The sort of thing which can go wrong is illustrated especially clearly by the seventh set of data in which cloud cover changes dramatically at $t_{hour} = 17$. The run which doesn't account for history assumes c_l has been zero since dawn and so predicts a much deeper boundary layer at $t_{hour} = 17$.

If solarimeter, sunshine, net radiation or heat flux data was used instead of cloud cover, it would be necessary to input data for previous hours since ADMS cannot estimate the history of $F_{\theta 0}$ etc without cloud data. (If the history isn't available, ADMS will estimate a neutral value of h which could be substantially in error. This is illustrated by the neutral values shown in figure 5.5.) However, since these types of measurements are more easily automated than cloud measurements, this should not be difficult to do.

A slight practical problem is that, if a history of met conditions is given, ADMS will calculate the dispersion for each hour, even though it may only be the last hour which is required. This could be avoided in principle by giving wind direction as missing except for the last hour of data – this should not prevent ADMS processing the met data but should stop it calculating dispersion except for the last hour. However, although this works in ADMS 1.5, the code fails on the first set of met data when this is tried with ADMS 2.02. With ADMS 2.02 one would have to calculate dispersion for each hour and ignore the unwanted cases.

(v) Properties above the boundary layer, N_u and $\Delta\theta$. Apart from use in connection with predicting h (discussed above), these are of no value except for buoyant releases and in any case would not be measurable by the 'type of simple measurements available at nuclear sites'. If one was considering a buoyant release, values could in principle be estimated from Met Office radiosonde ascents or NWP model output. However, plumes above the boundary layer and the question of whether a buoyant plume penetrates the boundary layer top in ADMS are treated rather simply and I judge that the use of such data instead of using the ADMS defaults is an unwarranted complication.

(vi) Wind direction standard deviation, σ_θ. Most (possibly all) sites have some means to estimate wind direction variability. However, because ADMS makes its own estimate of turbulent wind fluctuations and only accepts input data on the non-turbulent part, it is hard to use such information. This however is not meant to imply a lack of value in wind direction records (such as anemographs). A record of the actual wind direction and its evolution in time could be invaluable for post accident analysis, especially if the release continues for a long time.

(vii) Land-sea temperature difference, ΔT and precipitation rate P. There is not much to say about these variables except that they are required if the coastline module is to be used or wet deposition estimated (except possibly for a statistical wet deposition calculation). We have not investigated the extent to which using the coastline module can change the concentration predictions.

6 Statistics

The above has been written primarily with the aim of predicting concentrations on a given occasion. If statistics of concentration over many (hypothetical or real) events are required some additional comments are called for.

In ADMS statistical calculations are usually carried out by providing a number of typical met conditions with information on how frequently each type of condition occurs. If one characterizes the meteorology by $U(z)$, ϕ, c_l, t_{day}, t_{hour}, T_0 and r one will have to consider an enormous number of different typical met conditions to represent the climate – indeed it would probably be quicker to run ADMS with e.g. 10 years of hourly data rather than meteorological statistics. Also, it would not be possible for ADMS to take account of the history of the meteorology in calculating h. As a result, it is better to calculate $F_{\theta 0}$ and h before evaluating the met statistics. Then one can characterise the meteorology by just the four variables $U(z)$, ϕ, $F_{\theta 0}$ (or L_{MO}) and h. Unfortunately ADMS cannot be used easily to do the preprocessing (i.e. calculating $F_{\theta 0}$ and h), although it would be possible *in principle* to use code from ADMS or code used for preprocessing met data at the Met Office to do this. Alternatively the ADMS methods of calculating $F_{\theta 0}$ (see Appendix B) and h (not described in detail here) could be recoded.

7 Conclusions and recommendations

We have reviewed the meteorological input requirements of ADMS and the types of simple measurements available at nuclear sites, and have investigated how the requirements can be met from the available data. The main conclusions are as follows:

(i) There are not many options in deciding how the requirements can be met because of the limited available data.

(ii) Wind speed and direction data should be obtained from the anemometer closest to the height of release.

(iii) For most sites, the available data means that ADMS should be run with c_l, t_{day}, t_{hour}, T_0, r and α as inputs (in addition to U and ϕ). Small errors in cloud amount, such as that arising from use of cloud data from a nearby met site or due to observational error are not of crucial importance, although they can result in errors in plume spread of order 20%. Sensitivity of results to temperature is also not great and so temperature measurements, although useful, are not required to any great accuracy. If necessary, the ADMS default temperature could be used, although this could result in errors in plume spread of order 20%. Although an error of 20% in plume width is probably acceptable and, in any case, within the accuracy expected from dispersion models, it seems desirable to avoid an *extra* error of 20%, especially one which can be so easily avoided. Generally the ADMS default for r should be used. However r should be increased to 0.6 if there is substantial snow cover. This can make a large difference to ADMS's dispersion predictions and so an estimate of surface snow cover is important. The ADMS default value for α (i.e. 1.0) should normally be adopted although a smaller value of 0.5 would be appropriate to drought or urban conditions.

(iv) Use of solarimeters to measure incoming solar radiation K^+ is a possibility during the day. This would avoid the need to observe cloud amount. However K^+ data could not be used directly in ADMS and would require some 'pre-processing'. If such an approach was adopted during the day, then there is a problem over what to do at night. However the dependence on cloud cover at night is not great and so cloud amount from a nearby met site would be adequate.

(v) Ideally hourly measurements should be input into ADMS so that ADMS can use the meteorological history to estimate h during the day. If cloud cover is input, it will often be acceptable not to provide the met history, although on occasion it is possible to obtain large errors in boundary layer depth. If such occasional errors are unacceptable, the history should be given. If solarimeter data is used, it is essential to provide hourly data to obtain acceptable predictions of h.

It should be noted that the range of met conditions is very large and the above conclusions are necessarily based on a relatively small sample of conditions. However it seems probable that the conclusions are valid more generally.

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Appendix A: Using Pasquill stability data with ADMS

ADMS cannot use Pasquill stability data directly – indeed one of the prime motivations for ADMS was the desire to escape the theoretical inadequacies of the Pasquill stability concept as discussed in §2. In most situations there is also little reason to wish to do so – if one has enough information to calculate Pasquill stability then one also has enough to run ADMS directly. However, there is one type of situation where the use of Pasquill stability data may be useful, namely where one has archives or statistical analyses of historical met data which contain information on Pasquill stability but not on e.g. cloud cover.

Because different schemes for calculating Pasquill stability give somewhat different results it is important, in devising a scheme to use Pasquill stability, to take account of the particular method used to calculate the Pasquill stability originally. Here we consider the two standard methods used by the Met Office (Thomson and Tonkinson, 1992) – which is used in a given situation depends on whether intermediate categories such as 'A-B' are required and whether boundary layer depth estimates are also needed.

The first method deduces $\mathcal{P}$ from $F_{\theta 0}$ and wind speed U during the day (solar elevation > 0) and from U and 'modified' cloud amount c_{lm} at night (the 'modification' to the cloud amount is to account for the different effect of clouds at different heights and is based on Nielsen et al (1981)). The equations used are

$$\mathcal{P} = 7 - [2.26 + 0.019(\hat{U} - 5.6)^2][0.1\hat{F}_{\theta 0} + 2 + 0.4\hat{U}^{3/2}]^{[0.28 - 0.004(\hat{U}-2)^2]} \tag{2}$$

during the day and

$$\mathcal{P} = 3.6 + \frac{120 - 13.3c_{lm}}{27 - 2c_{lm}} \exp\left(-\frac{3}{8}\hat{U}\right) \tag{3}$$

at night. Here $\hat{U} = \min(U, 8)$, $\hat{F}_{\theta 0} = \max(F_{\theta 0}, 0)$ with U, $F_{\theta 0}$ and c_{lm} measured in m/s, W/m^2 and oktas respectively. The daytime formula is an expression fitted by Farmer (1984) to the curves given by Smith (1973) (see also Pasquill and Smith (1983, p337)), while the nighttime formula is a proposal of Smith (1983 – unpublished) which is reported in Farmer (1984). Both formulae are plotted in figure A1. $F_{\theta 0}$ is determined using a scheme based on Berkowicz and Prahm (1982) and Nielsen et al (1981). $\mathcal{P}$ is then converted to a letter as follows

numeric $\mathcal{P}$	alphabetic $\mathcal{P}$
$\mathcal{P} < 1$	A
$1 \le \mathcal{P} < 2$	B
$2 \le \mathcal{P} < 3$	C
$3 \le \mathcal{P} < 4$	D
$4 \le \mathcal{P} < 5$	E
$5 \le \mathcal{P} < 6$	F
$6 \le \mathcal{P}$	G

The second method is closer conceptually to Pasquill's original method and is described in Farmer (1984). It uses U (in knots) and total cloud amount c_l (in oktas). During the day the incoming solar radiation K^+ (in W/m^2) is estimated as $880sf(c_l)$ where s is the sine of the solar elevation and f takes the following values

c_l :	0	1	2	3	4	5	6	7	8
$f(c_l)$:	1.07	0.89	0.81	0.76	0.72	0.67	0.59	0.45	0.23

(see Smith (1973)). If $s < 5/44$ the Pasquill Stability is taken to be category D (this is intended to account for times within about 1 hour of dawn and dusk). Otherwise the Pasquill Stability is estimated using the following table:

Table A1	$c_l = 8$	$c_l < 8$		
		$K^+ < 300$	$300 \le K^+ < 600$	$600 \le K^+$
$U \le 3$	C	B	A-B	A
$3 < U \le 5$	C	C	B	A-B
$5 < U \le 9$	C	C	B-C	B
$9 < U \le 12$	D	D	C-D	C
$12 < U$	D	D	D	C

During the night $\mathcal{P}$ is estimated directly from U and c_l using the following table:

Table A2	$c_l = 0, 1$	$c_l = 2, 3$	$c_l = 4, 5, 6, 7$	$c_l = 8$
$U \le 1$	G	F	F	D
$1 < U \le 3$	F	F	F	D
$3 < U \le 5$	F	F	E	D
$5 < U \le 9$	E	E	D	D
$9 < U$	D	D	D	D

To illustrate the variation which is possible between different Pasquill schemes, a comparison of the frequency of occurrence of the various categories as calculated by the two schemes is shown in figure A2. The data is from Belfast (Aldergrove Airport) and, to make the comparison easier, the intermediate categories in the second method have been split evenly between adjacent categories. Although both cases are dominated by a 60 to 70% occurrence of category D, the frequency distribution of the other categories show significant differences. This illustrates the difficulty associated with Pasquill stability discussed in §2 – because one cannot measure Pasquill stability one cannot make an objective assessment of which scheme is better.

The simplest way of using such values of $\mathcal{P}$ in ADMS is to use them to estimate $F_{\theta 0}$ which can then be input to ADMS (Thomson and Tonkinson, 1992). ADMS could then be run using data on wind speed, direction, $\mathcal{P}$ and boundary layer depth h (and rainfall if wet deposition estimates are required and both sea surface temperature and near surface temperature over land or the difference between them if the coastline module is to be used), or even on just wind speed, direction and $\mathcal{P}$ as ADMS will make an estimate of h if this isn't provided.

For $\mathcal{P}$ calculated with the first scheme, it is relatively easy to estimate $F_{\theta 0}$. First $\mathcal{P}$ is converted back to a numeric value as follows

alphabetic $\mathcal{P}$	numeric $\mathcal{P}$
A	0.5
B	1.5
C	2.5
D	3.5
E	4.5
F	5.5
G	6.5

For unstable conditions (i.e. for $\mathcal{P} = A$, B or C or, in the numeric equivalent, $\mathcal{P} = 0.5$, 1.5 or 2.5) we can assume equation (2) has been used. This is easily inverted to give

$$F_{\theta 0} = \min[300, 10(r - 2 - 0.4\hat{U}^{3/2})] \tag{4}$$

where

$$r = \left(\frac{7 - \mathcal{P}}{2.26 + 0.019(\hat{U} - 5.6)^2}\right)^{1/[0.28 - 0.004(\hat{U}-2)^2]}.$$

Because accuracy has been lost in converting $\mathcal{P}$ to a letter and then back to a number it has been necessary to limit $F_{\theta 0}$ to $300W/m^2$ in (4). Without the limitation on $F_{\theta 0}$ it is possible to obtain some unrealistically large heat fluxes. In near-neutral conditions (i.e. $\mathcal{P} = D$) we recommend taking $F_{\theta 0} = 0$. Finally for stable conditions (i.e. $\mathcal{P} = E$, F or G or, in the numeric equivalent, $\mathcal{P} = 4.5$, 5.5 or 6.5) we can assume equation (3) has been used. In this case (3) can be easily inverted to give c_{lm} using

$$c_{lm} = \begin{cases} 0 & r \geq 4.44 \\ \frac{120-27r}{13.3-2r} & 1.24 < r < 4.44 \\ 8 & r < 1.24 \end{cases}$$

where

$$r = (\mathcal{P} - 3.6)\exp\left(\frac{3}{8}\hat{U}\right).$$

As in (4) we have applied corrections for the loss of accuracy in $\mathcal{P}$. $F_{\theta 0}$ can then be estimated using any of the nighttime schemes which express $F_{\theta 0}$ in terms of cloud amount and wind speed. ADMS has such a scheme (due to Holtslag and van Ulden (1982) and described in Appendix B) built in and the user could simply supply U and c_{lm} to the model together with a time of day and year (these times need not be correct but must be such as to allow the model to deduce that it is nighttime).

For the second scheme a different approach must be adopted. In unstable conditions, table A1 can be approximately inverted (ignoring the $c_l = 8$ column) to give K^+ in terms of U and $\mathcal{P}$. $F_{\theta 0}$ can then be estimated using Smith's (1973) formula, $F_{\theta 0} = 0.4(K^+ - 100)$. The result is the following table of $F_{\theta 0}$ values:

Table A3	A	A-B	B	B-C	C	C-D
$U \leq 3$	260	140	20	(15)	10	(5)
$3 < U \leq 5$	(300)	260	140	(80)	20	(10)
$5 < U \leq 9$	(300)	(300)	260	140	20	(10)
$9 < U \leq 12$	(300)	(300)	(300)	(300)	260	140
$12 < U$	(300)	(300)	(300)	(300)	260	(140)

Here we have assumed $K^+ = 150$, $K^+ = 450$ and $K^+ = 750$ for the classes $K^+ < 300$, $300 \leq K^+ < 600$ and $600 \leq K^+$ and have interpolated and extrapolated the $F_{\theta 0}$ values, subject to $F_{\theta 0}$ not exceeding 300W/m^2. Of course some combinations of U and $\mathcal{P}$ should not occur – these are indicated by parentheses. In near-neutral conditions we adopt $F_{\theta 0} = 0$, as for the first scheme. Finally in stable conditions table A2 can be approximately inverted to give c_l:

Table A4	E	F	G
$U \leq 1$	(8)	4.5	0.5
$1 < U \leq 3$	(8)	3.5	(0)
$3 < U \leq 5$	5.5	1.5	(0)
$5 < U \leq 9$	1.5	(0)	(0)
$9 < U$	(0)	(0)	(0)

As for $F_{\theta 0}$, we have extrapolated the c_l values, subject to c_l lying in [0,8] and have indicated combinations of c_l and $\mathcal{P}$ which should not occur by parentheses. As for the first scheme, $F_{\theta 0}$ can then be estimated using any of the nighttime schemes which express $F_{\theta 0}$ in terms of cloud amount and wind speed.

It should be emphasised that, as a consequence of the approximations involved in inverting the calculation of $\mathcal{P}$, results are likely to be inferior to those obtained directly by running ADMS on the raw data without using $\mathcal{P}$ as an intermediate step. The first cause of differences involves the mathematical errors arising from the fact that a given Pasquill stability category covers a range of conditions and so the schemes do not possess a unique 'inverse'. There will also be differences resulting from different assumptions about the physics. There are two instances of this, both relating to unstable daytime conditions. The first scheme for calculating $\mathcal{P}$ uses $F_{\theta 0}$ as an intermediate variable and calculates $F_{\theta 0}$ in a different way to that in ADMS. The scheme proposed to calculate $F_{\theta 0}$ from $\mathcal{P}$ simply attempts to invert this calculation and so the differences in the $F_{\theta 0}$ schemes will cause further differences from the results that would be obtained by running ADMS on the raw data. It is not practical to do more than this and, in any case, the $F_{\theta 0}$ scheme used (Berkowicz and Prahm (1982), Nielsen et al (1981)) is of comparable accuracy to the ADMS scheme (Galinski and Thomson 1995). Similarly the second scheme uses a different approach from that in ADMS for estimating K^+ and the scheme proposed for estimating $F_{\theta 0}$ from K^+ (namely Smith's (1973) formula, $F_{\theta 0} = 0.4(K^+ - 100)$) is also rather different from that in ADMS. Again it is not practical to adopt an approach closer to that used by ADMS.

Appendix B: Calculating $F_{\theta 0}$ and L_{MO} from routinely available data

This appendix describes the method used in ADMS to calculate the heat flux $F_{\theta 0}$ and Monin-Obukhov length L_{MO} from routinely available data. Although heat flux $F_{\theta 0}$ is not a 'routinely available quantity' we will discuss first the simpler problem of determining L_{MO} from $F_{\theta 0}$, and then discuss the question of what to do if $F_{\theta 0}$ is not available.

Equation (1) above defines L_{MO}. If $F_{\theta 0}$ is known, the only remaining unknowns are T_0 and u_*. T_0 is the absolute temperature and so doesn't need to be known very accurately – if it's not measured taking $T_0 = 288.15\text{K}$ ($= 15^o\text{C}$) would be an adequate approximation. That leaves u_* as an unknown. Conventional surface layer theory allows us to relate u_* to U (at some height z), L_{MO} and z_0 as follows:

$$U(z) = \frac{u_*}{k}\left(\log\left(\frac{z+z_0}{z_0}\right) + \psi\left(\frac{z+z_0}{L_{MO}}\right) - \psi\left(\frac{z_0}{L_{MO}}\right)\right) \tag{5}$$

where ψ is a semi-empirical function which will be discussed further below (Panofsky and Dutton, 1984). For given ψ, (1) and (5) can be solved simultaneously for u_* and L_{MO}.

There are two complications in solving these equations. The first concerns the fact that in stable conditions the form of ψ is such that the equations may, for certain values of z_0, $U(z)$, $F_{\theta 0}$ and T_0 have no solution or more than one solution. This can be understood by considering what happens as u_* varies for fixed z_0, $F_{\theta 0}$ and T_0. For large values of u_*, the flow is close to neutral and

$$U(z) \simeq \frac{u_*}{k}\log\left(\frac{z+z_0}{z_0}\right).$$

As u_* decreases, the stability (e.g. as measured by $1/L_{MO}$) increases and $U(z)$ at first decreases. However as the stability continues to increase, the effect of the ψ function is to allow more shear $\frac{\partial U}{\partial z}$ for a given stress u_*^2 (in more stable flows momentum transfer is less effective due to reduced turbulence) and eventually $U(z)$ increases after reaching a minimum value. If the measured $U(z)$ is less than the minimum the equations have no solution while for larger values of $U(z)$ there will be at least two solutions. In the case of two solutions, the larger value of u_* corresponds to more neutral conditions, higher wind speeds near the ground, and lower wind shear above the lowest layers, while the smaller value corresponds to more stable conditions, lower wind speeds near the ground, but more wind shear (see figure B1). If there is no solution this means that, according to the model wind profile (5), the data used do not represent a possible state of the atmosphere and that, for the given values of z_0, $U(z)$ and T_0, the value of $|F_{\theta 0}|$ exceeds the maximum amount of heat which the atmosphere is capable of transporting. (The caveat 'according to the model' is meant to indicate, not that there is any serious doubt about the qualitative behaviour of the model, but that the precise boundary of the possible states may not be correct in the model). Physically, the fact that $|F_{\theta 0}|$ is limited can be understood by noting that, if $|F_{\theta 0}|$ is increased, the stability increases and the turbulence becomes less able to transport heat. The simplest procedure in this case is to replace $|F_{\theta 0}|$ by the largest value for which the equations do have a solution. If there are two or more possible solutions the best procedure is to choose the most neutral one (i.e. the one with the largest u_*) since otherwise one could not treat near neutral conditions correctly.

Unfortunately however there is no physical basis for deciding which solution is correct without making more sophisticated measurements such as direct measurements of u_*. By choosing the most neutral solution one is biasing the results a little and this effect becomes worse the larger is the height of the wind measurement. For $\psi(\zeta) = 5\zeta$, a widely used parametrization for stable conditions (see e.g. Panofsky and Dutton (1984)), the minimum possible value of L_{MO} is $10z/\log((z+z_0)/z_0)$, and so the minimum value of L_{MO} is more restrictive if measurements are made at higher heights. For the ADMS form of ψ, it is not possible to evaluate the minimum value of L_{MO} analytically; however the minimum value is generally less restrictive than that for $\psi(\zeta) = 5\zeta$.

The second complication concerns finding a suitable solution technique. This can only be considered in relation to specific assumptions on ψ. In ADMS ψ is given by

$$\psi(\zeta) = \begin{cases} a\zeta + b(\zeta - c/d)\exp(-d\zeta) + bc/d & \text{if } 1/L_{MO} \geq 0 \\ 2\tan^{-1} x - \log((1+x)^2(1+x^2)) & \text{otherwise} \end{cases}$$

where $a = 0.7$, $b = 0.75$, $c = 5$, $d = 0.35$ and $x = (1 - 16\zeta)^{1/4}$ (Holtslag and de Bruin, 1988; Dyer and Hicks, 1970; Benoit, 1977). For this form of ψ it can be shown that, in stable conditions ($F_{\theta 0} \leq 0$), the following iteration technique always works.

- (i) Calculate u_* from (5) as if conditions are neutral (i.e. $1/L_{MO} = 0$).
- (ii) Calculate L_{MO} from (1).
- (iii) Calculate u_* from (5) using the value of L_{MO} from (ii).
- (iv) Repeat (ii) and (iii) until the solution converges.

If, at the end of stage (ii), the most recently calculated values of u_* and L_{MO} ever reach values which satisfy

$$(a - b\exp(-(c+z)))\frac{z}{L_{MO}} > \frac{1}{2}\log\left(\frac{z+z_0}{z_0}\right)$$

and

$$(a - b\exp(-(c+z)))\frac{z}{L_{MO}} > \frac{kU(z)}{u_*} - \log\left(\frac{z+z_0}{z_0}\right),$$

then there is no solution and $|F_{\theta 0}|$ needs to be reduced. If there are two or more solutions, the procedure always converges to the most neutral solution. For convective conditions ($F_{\theta 0} > 0$), the following procedure can be shown to work.

- (i) Calculate u_* from (5) as if conditions are neutral (i.e. $1/L_{MO} = 0$). Call this value u_{*L} – it is a lower limit on u_*.
- (ii) Calculate L_{MO} from (1) using u_{*L}.
- (iii) Calculate u_* from (5) using the value of L_{MO} from (ii). Call this value u_{*U} – it is an upper limit on u_*.

- (iv) Calculate $u_{*M} = (u_{*L} + u_{*U})/2$, calculate L_{MO} from (1) using u_{*M} and calculate $U(z)$ from (5) using u_{*M} and L_{MO}. If the calculated $U(z)$ is larger than the measured value then u_{*M} is an upper limit to u_* - set u_{*U} equal to u_{*M}. If not then u_{*M} is a lower limit to u_* - set u_{*L} equal to u_{*M}.
- (v) Repeat (iv) until the difference between and u_{*L} and u_{*U} becomes small enough.

We now consider the situation where $F_{\theta 0}$ is not known and has to be estimated. First ADMS calculates the sine s of the solar elevation using

$$s = \sin(2\pi S_{lat}/360)\sin(2\pi A/360) + \cos(2\pi S_{lat}/360)\cos(2\pi A/360)\cos(2\pi(t_{hour} - 12)/24)$$

where A, the solar declination in degrees, equals $23.45\sin(2\pi(t_{day} + 284)/365)$ and S_{lat} is the latitude, also in degrees (Ratto 1988). We now consider two cases corresponding to night and day ($s \leq 0$ and $s > 0$). At night $F_{\theta 0}$ is calculated from

$$F_{\theta 0} = -\rho_a c_p u_* \theta_*$$

with

$$\theta_* = 0.09(1 - 0.5(c_l/8)^2)$$

where θ_* here is measured in °C (Holtslag and van Ulden, 1982; van Ulden and Holtslag, 1983). This depends on u_* and so $F_{\theta 0}$ has to be found in conjunction with u_* and L_{MO} using iterative methods similar to those described above. As above there are problems with no solutions or multiple solutions occurring. In the first case θ_* should be reduced to get a solution while in the second the most neutral solution should be adopted.

For daytime conditions, the steps in the procedure follow the path of energy in the atmosphere. First incoming solar radiation K^+ is calculated from solar elevation and cloud cover, then net radiation Q_* is calculated from K^+ by accounting for the albedo and long wave radiation, and finally $F_{\theta 0}$ is calculated from Q_* by accounting for the ground heat flux and the partition of available energy between evaporation of water and the flux of sensible heat. The equations used follow Holtslag and van Ulden (1983) and are given by

$$K^+ = (990s - 30)(1 - 0.75(c_l/8)^{3.4}),$$

$$Q_* = \frac{(1-r)K^+ + 5.31 \times 10^{-13}T_0^6 - 5.67 \times 10^{-8}T_0^4 + 60(c_l/8)}{1.12},$$

and

$$F_{\theta 0} = \frac{(1-\alpha)S + 1}{S + 1}0.9Q_* - 20\alpha$$

with $S = \exp(0.055(T_0 - 279))$ (van Ulden and Holtslag 1985). Here T_0 is measured in Kelvin and K^+, Q_* and $F_{\theta 0}$ in Wm^{-2}. If values of T_0, α and r are not input, defaults of 288.15K, 1.0 and 0.23 are used. The value of T_0 is somewhat more important here than in the formula for L_{MO} and so adopting a default value is less desirable. α represents the availability of water for evaporation with 1.0 (the ADMS default) recommended by Holtslag and van Ulden (1983) for moist grassland. Smaller values may be appropriate for urban areas or during droughts (for example Holtslag and van Ulden (1983) found that $\alpha = 0.45$ was appropriate to the dry grassland in the Project Prairie Grass experiment). r is the albedo of the ground and 0.23 represents a typical mid-latitude value. Near dawn and dusk the heat flux from the scheme can become negative. To help ensure a smooth transition to the night time scheme, $F_{\theta 0}$ is replaced by the value calculated using the night time scheme if this is less negative.

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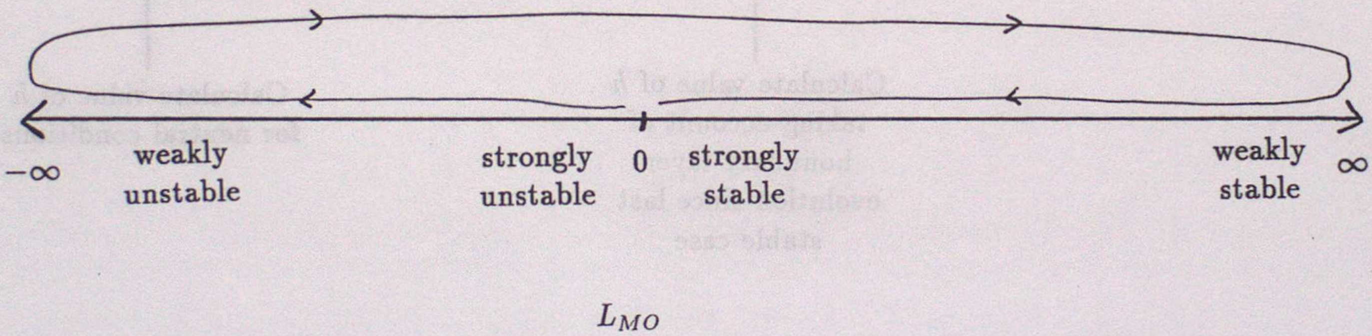


Figure 2.1: Illustration of how L_{MO} evolves as stability changes from strongly unstable to strongly stable.

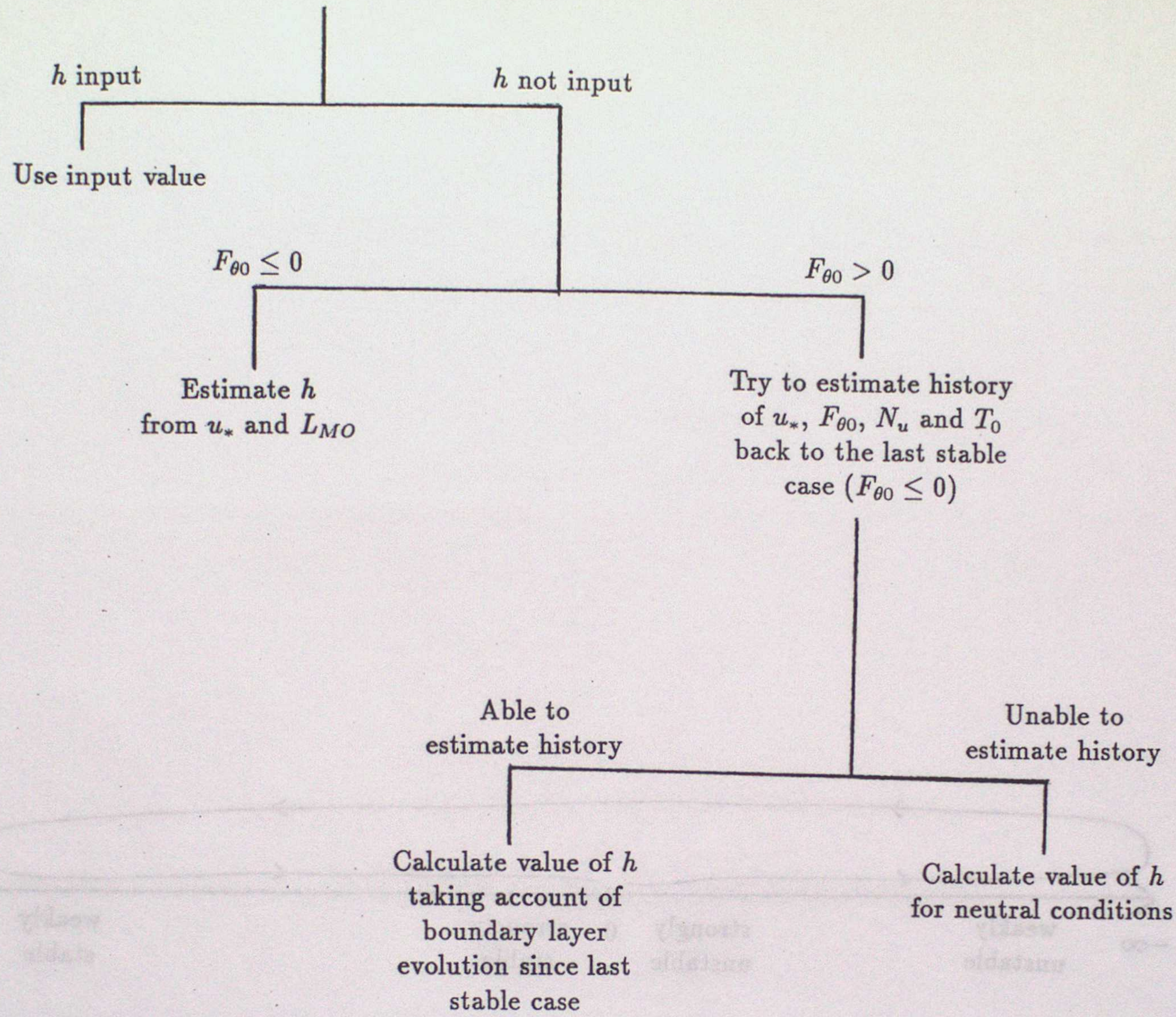


Figure 3.1: Methods used by ADMS to estimate boundary layer depth.

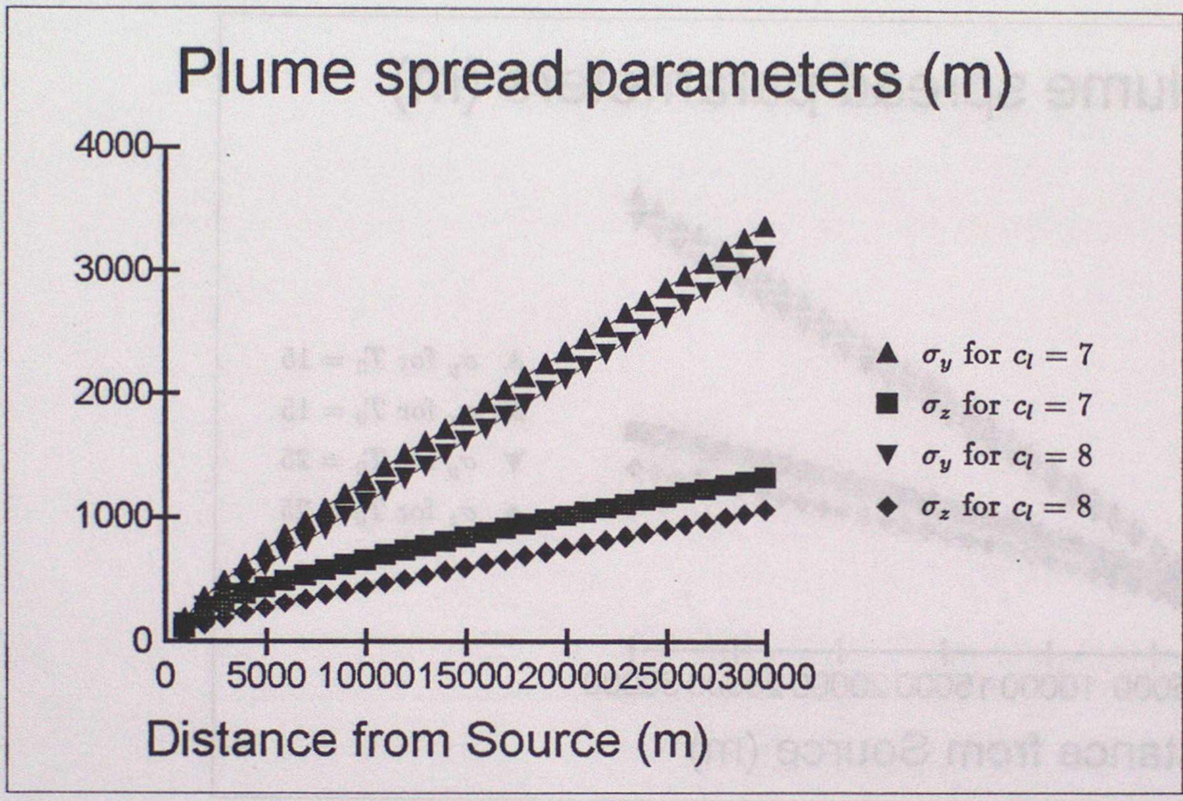


Figure 5.1: σ_y and σ_z for $c_l = 7$ and 8. For details of source and other met parameters see text.

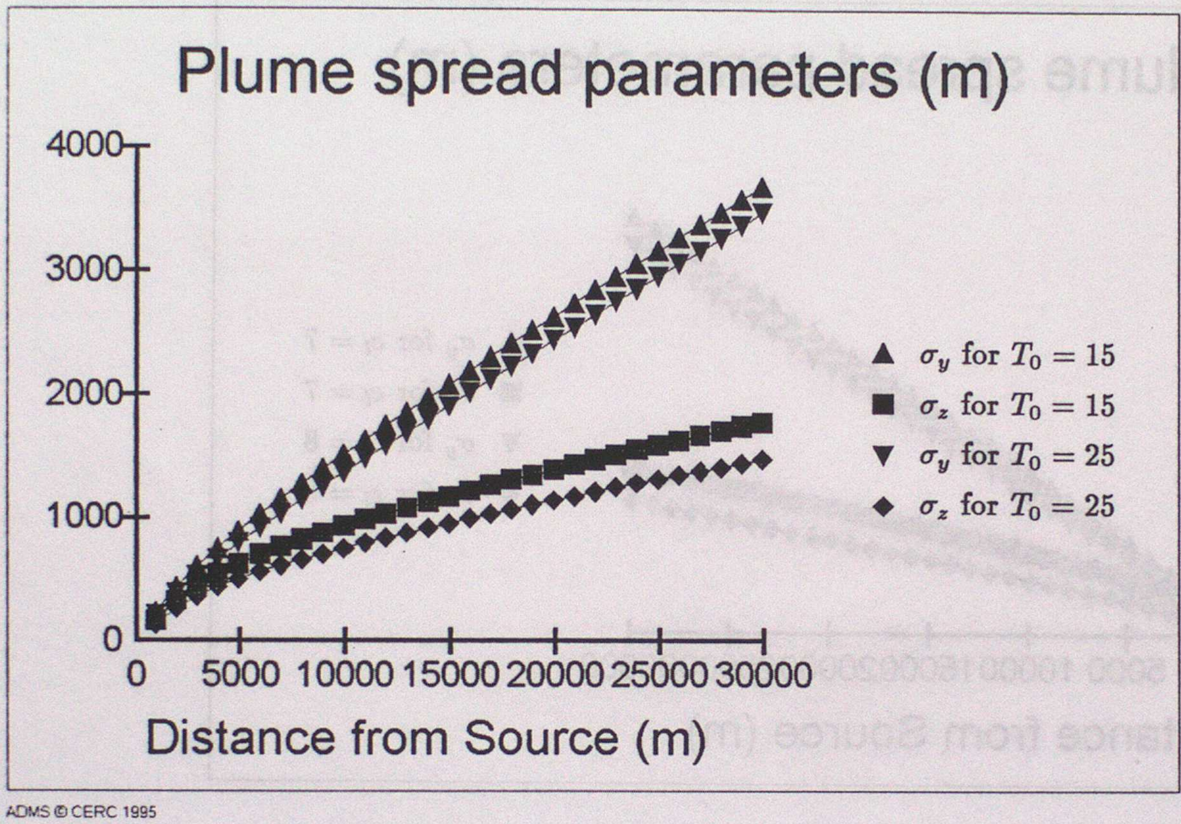


Figure 5.2: σ_y and σ_z for $T_0 = 15$ and 25. For details of source and other met parameters see text.

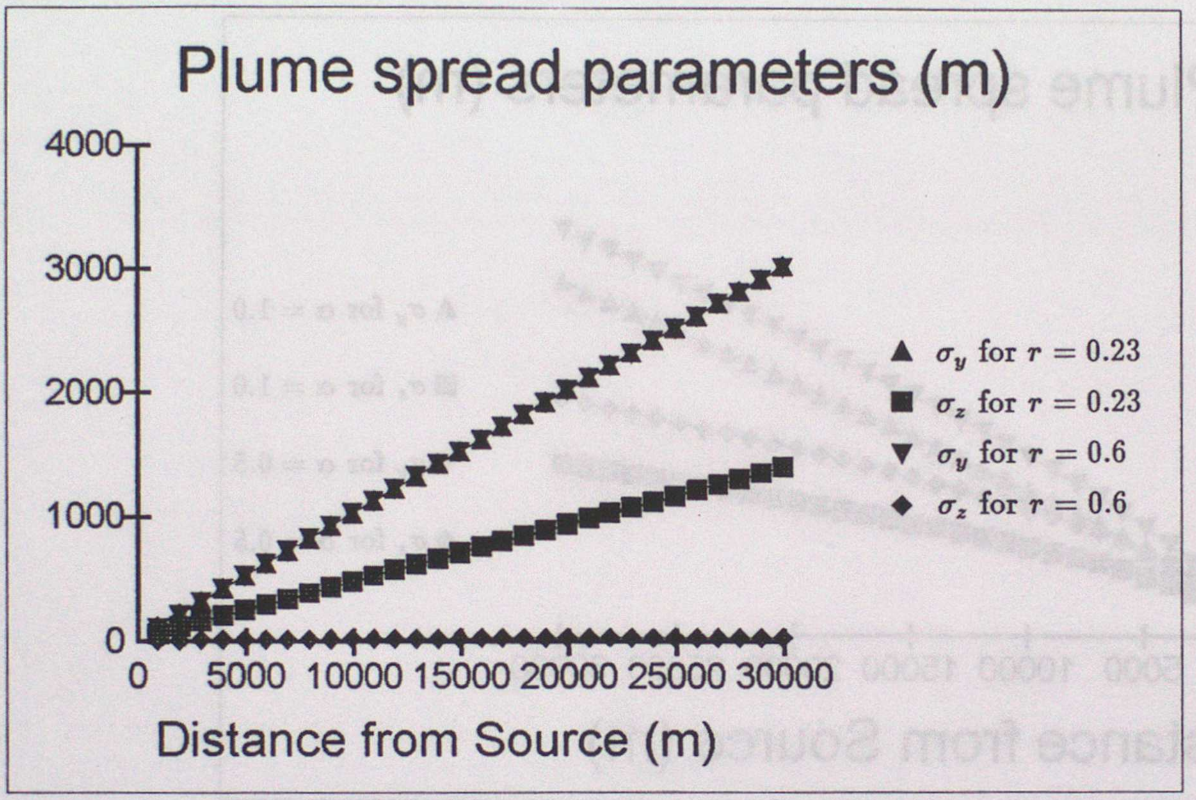


Figure 5.3: σ_y and σ_z for $r = 0.23$ and 0.6. For details of source and other met parameters see text.

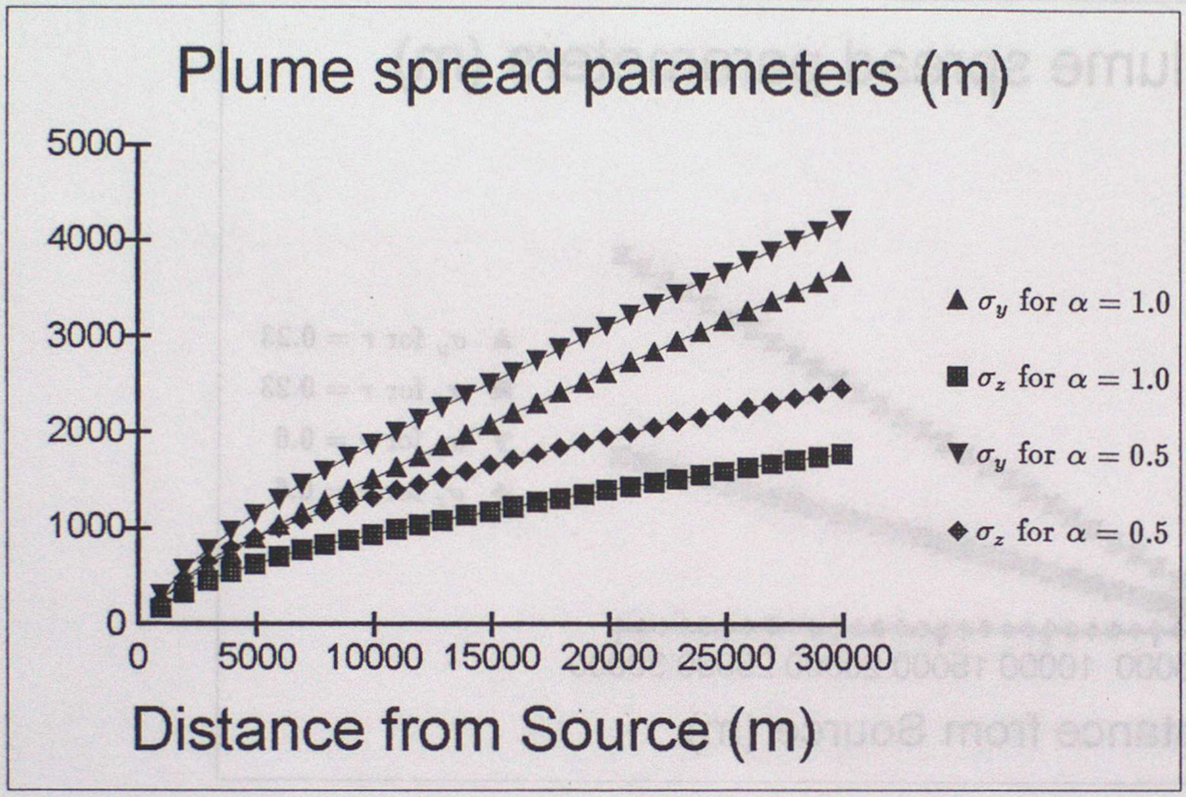


Figure 5.4: σ_y and σ_z for $\alpha = 1.0$ and 0.5. For details of source and other met parameters see text.

(a)

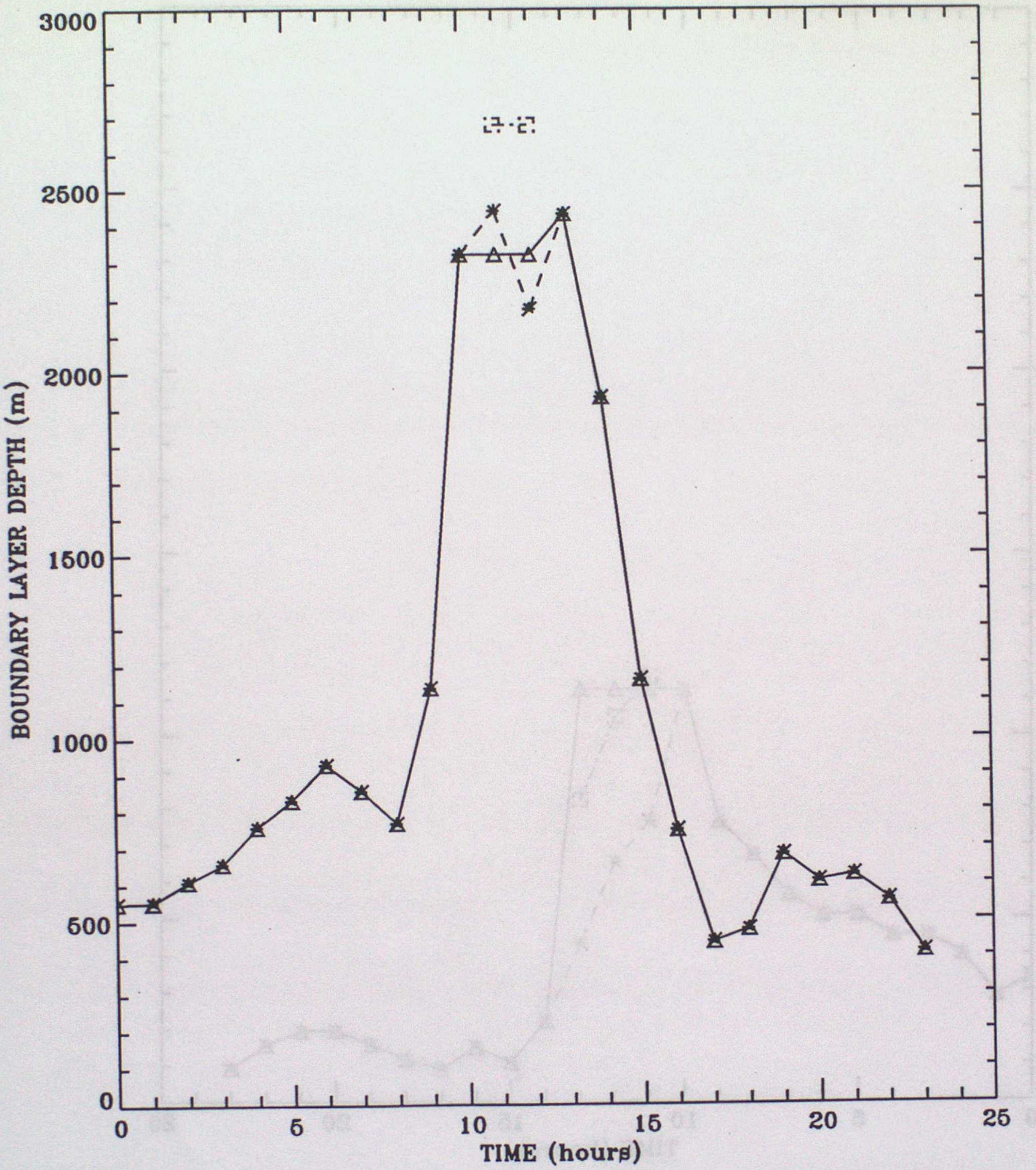


Figure 5.5: Estimates of boundary layer depth from ADMS. Figures (a) to (f) show results using data from Wattisham for 1/1/85, 1/3/85, 1/5/85, 1/7/85, 1/9/85 and 1/11/85 respectively, while figure (g) shows results for 1/7/85 but with the cloud amount set to 8 oktas up to $t_{hour} = 16$ and set to zero at later times. The solid line shows ADMS predictions which take account of the history of the meteorology, the dashed line shows results ignoring the history and the dot-dash line shows neutral boundary layer depth estimates. The neutral estimates are given for cases with $F_{\theta 0} > 0$ only.

(b)

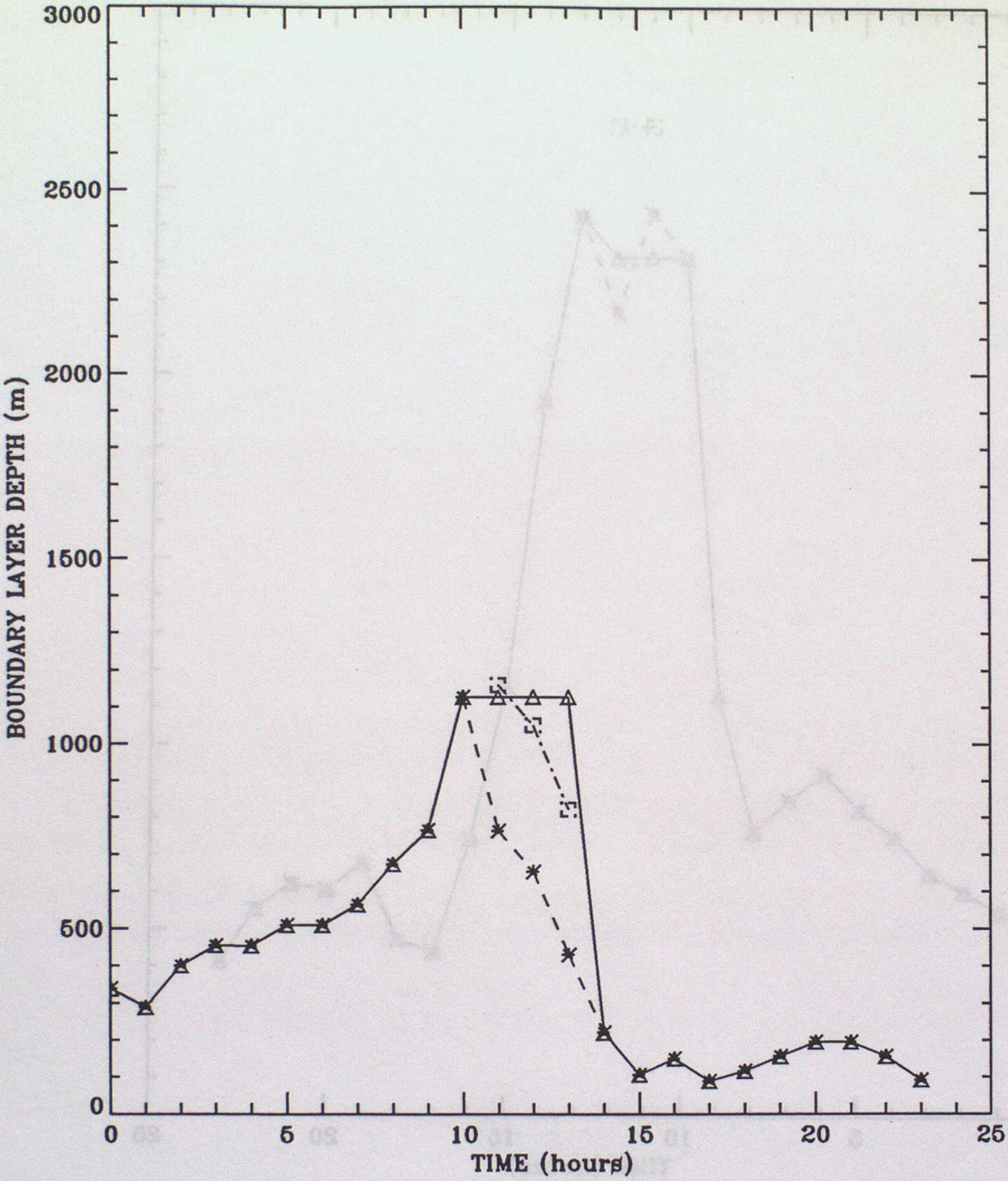


Figure 5.5 continued.

(c)

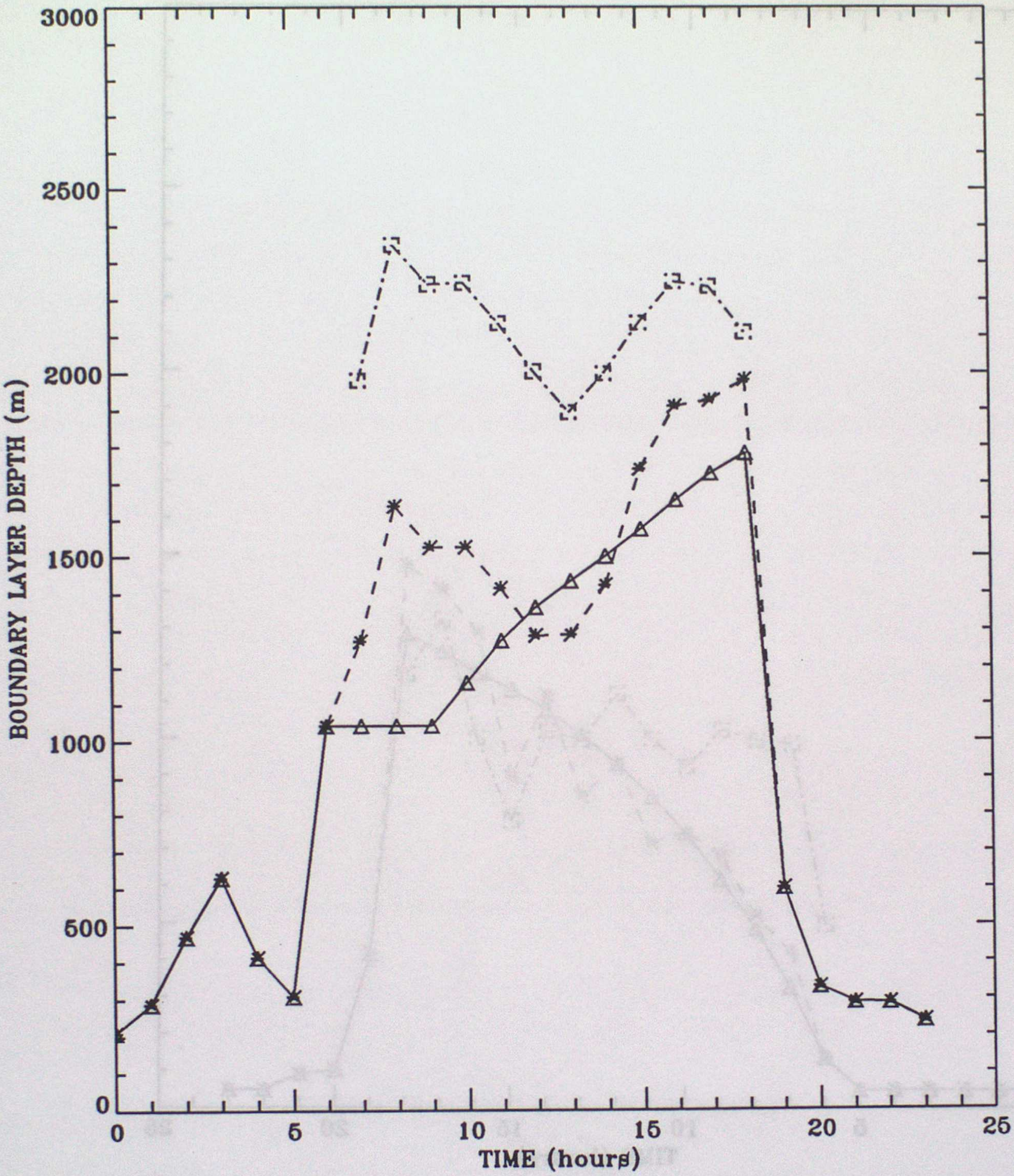


Figure 5.5 continued.

(d)

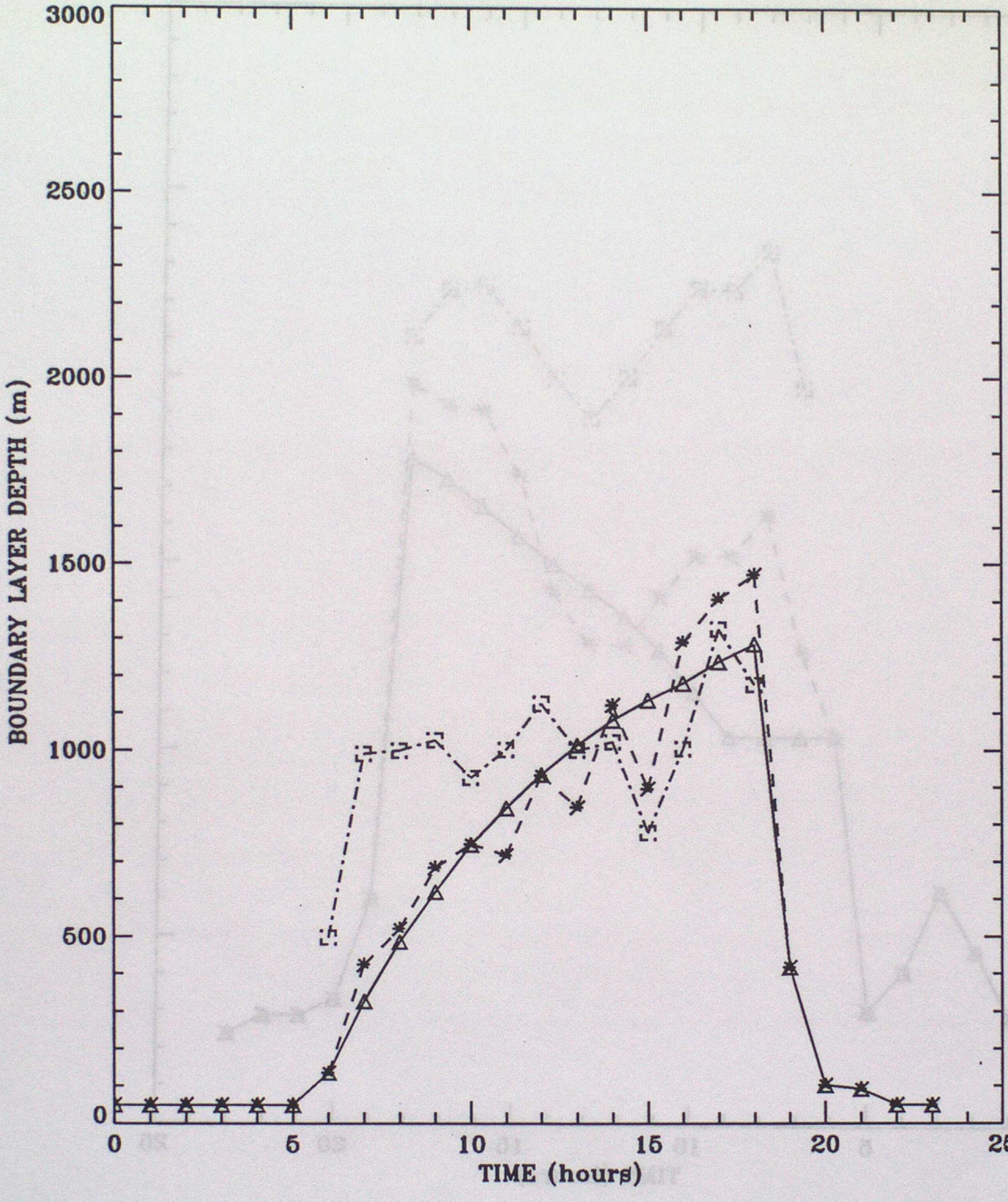


Figure 5.5 continued.

(e)

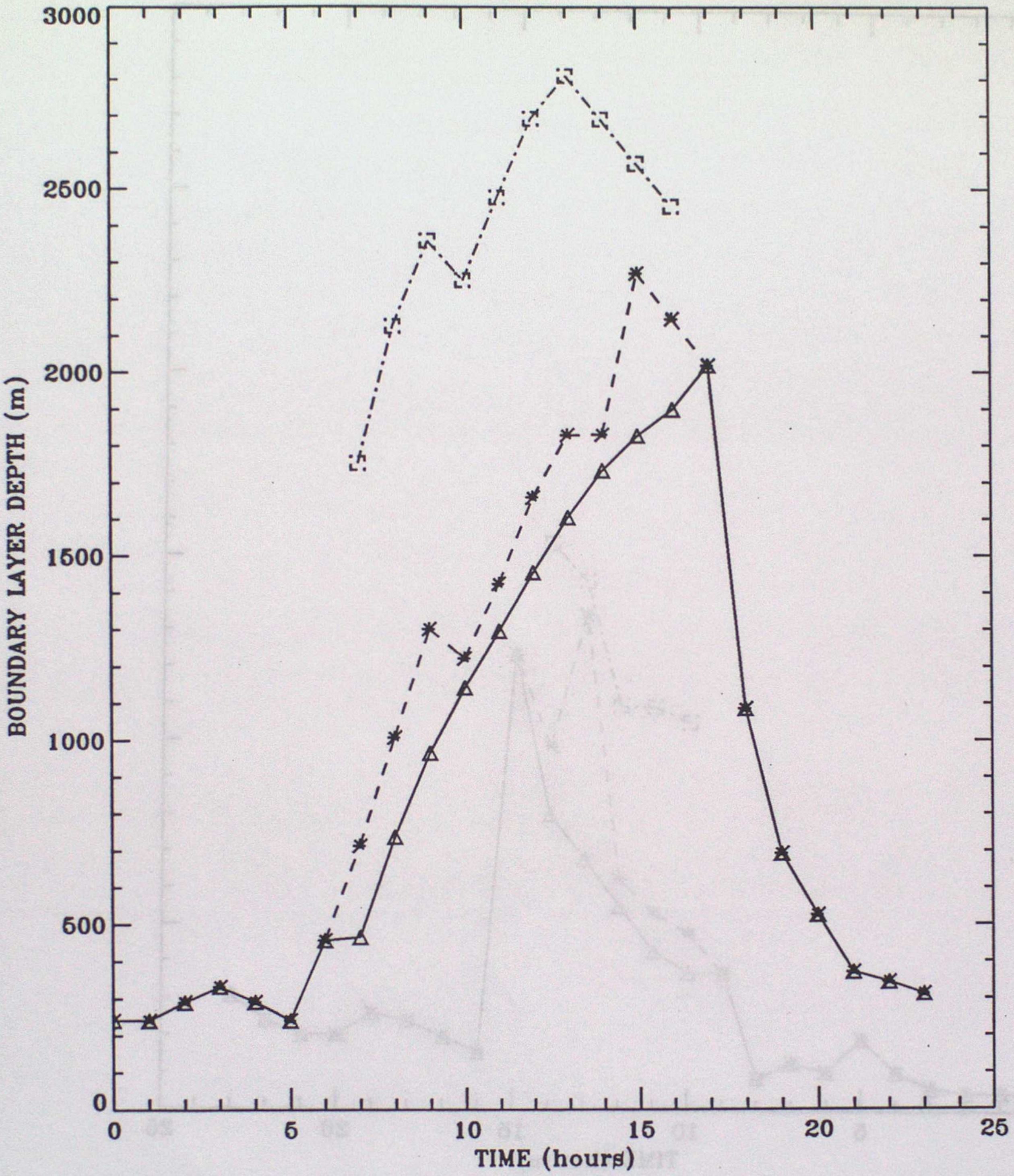


Figure 5.5 continued.

(f)

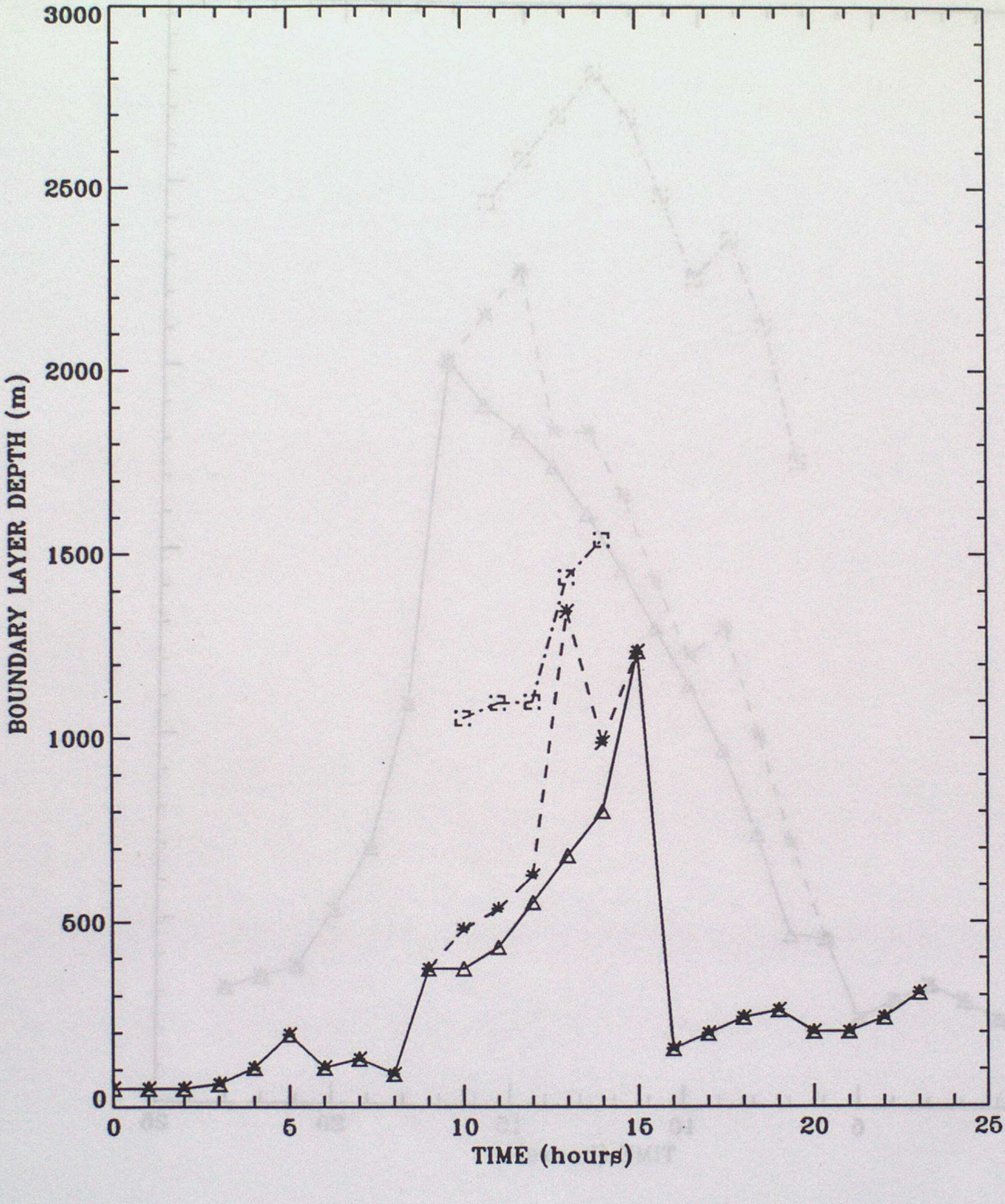


Figure 5.5 continued.

(g)

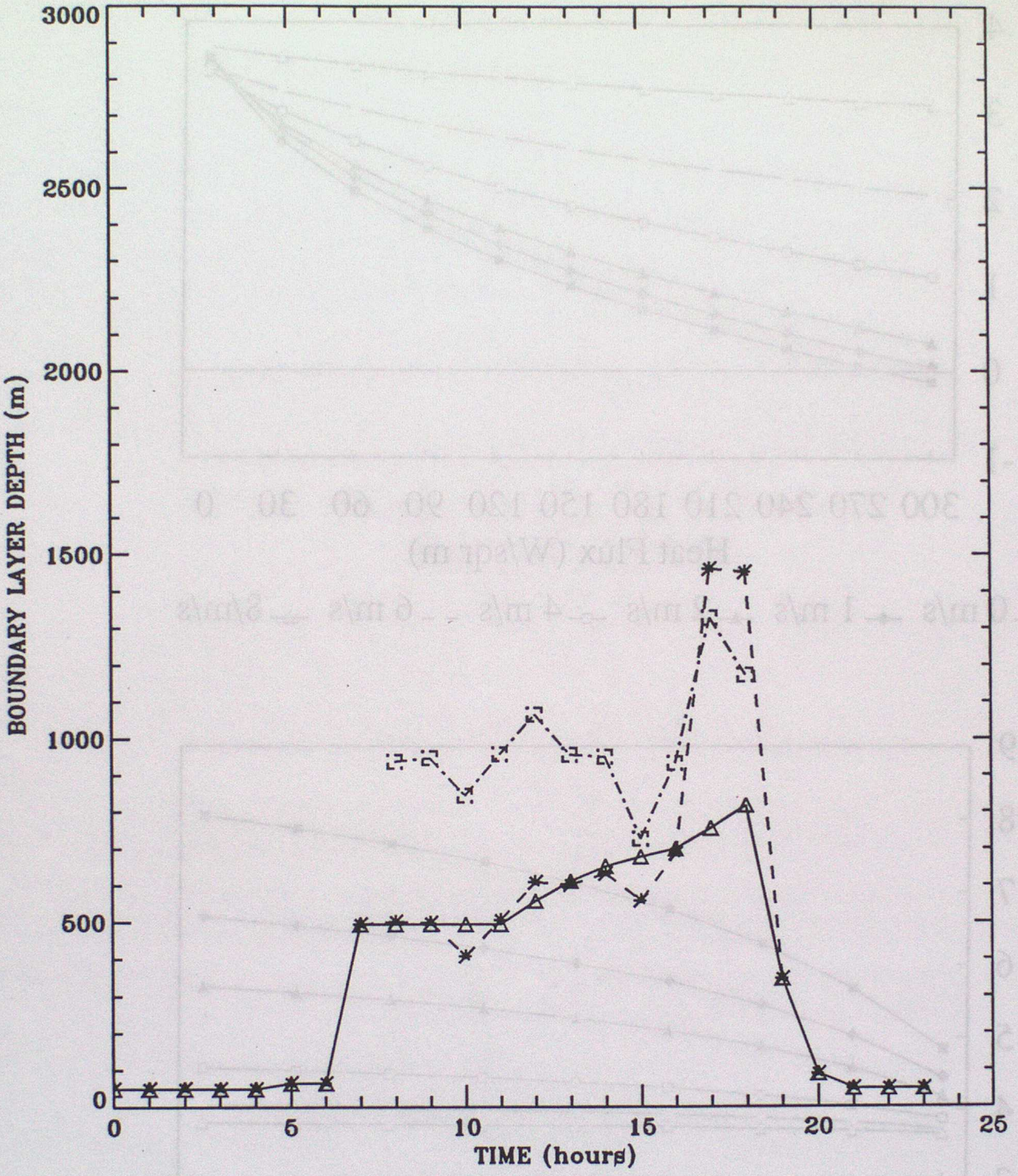
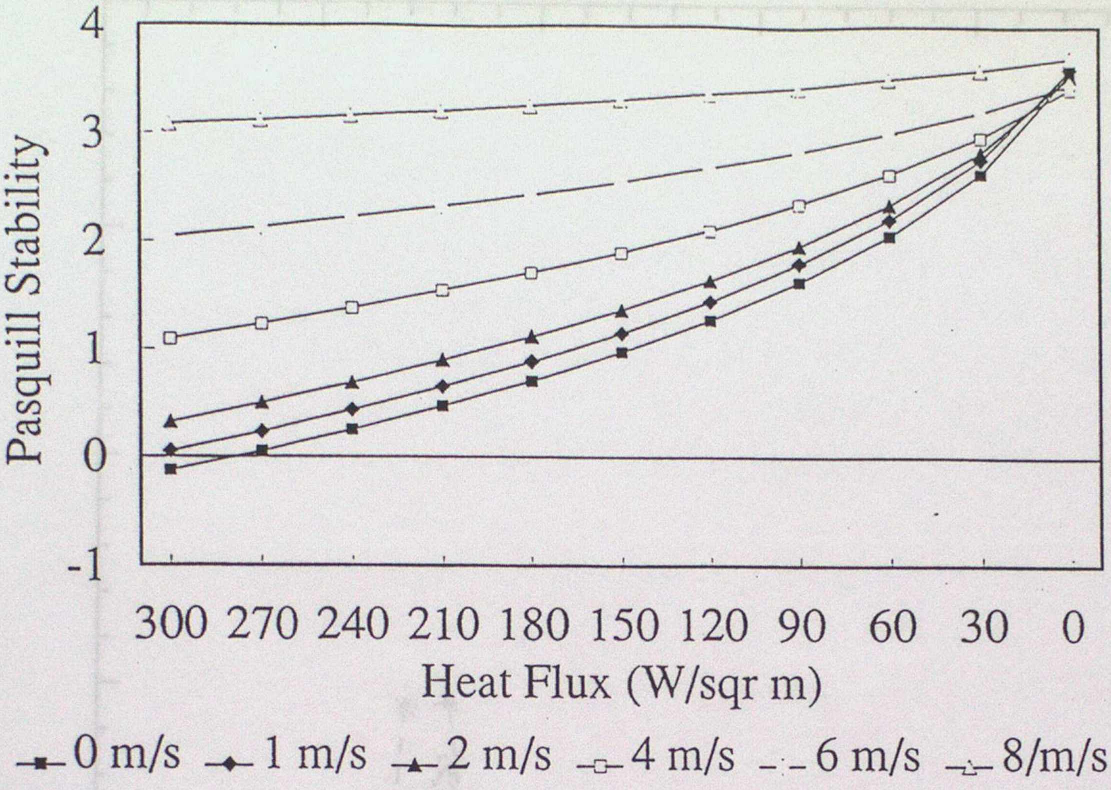


Figure 5.5 continued.

(a)



(b)

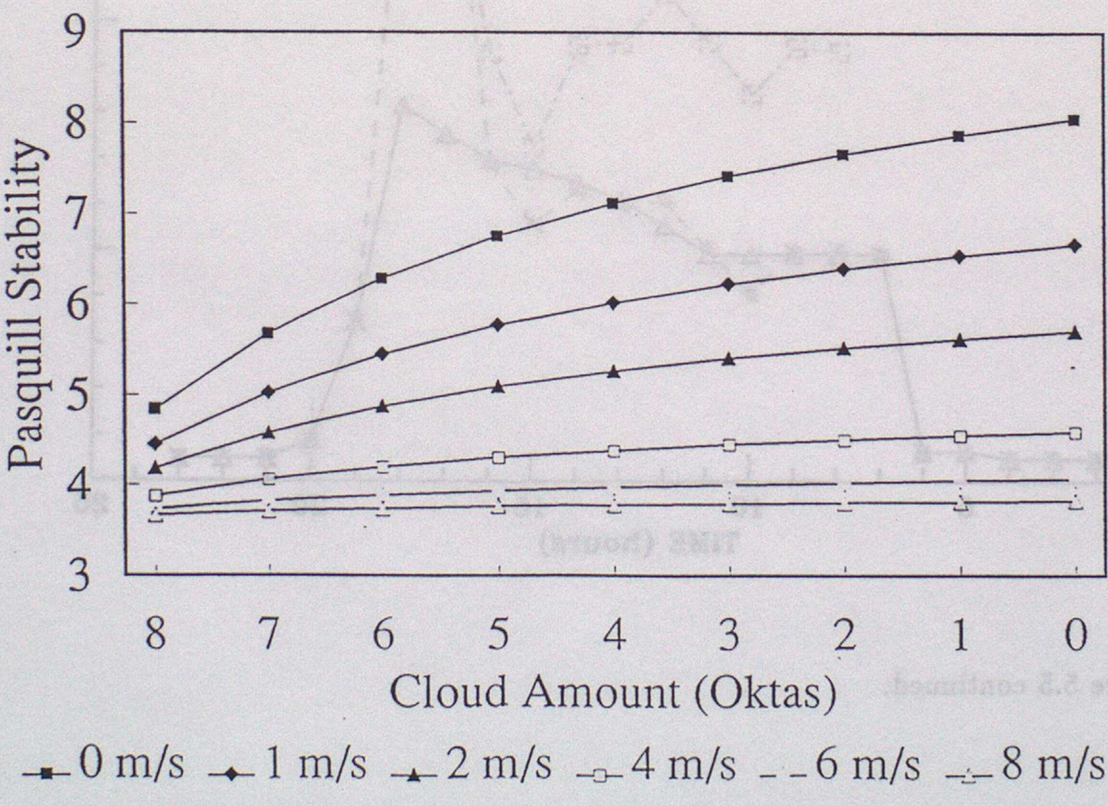


Figure A1: Pasquill Stability $\mathcal{P}$ as a function of wind speed and surface sensible heat flux during the day (figure A1(a)) and of wind speed and cloud amount at night (figure A1(b)).

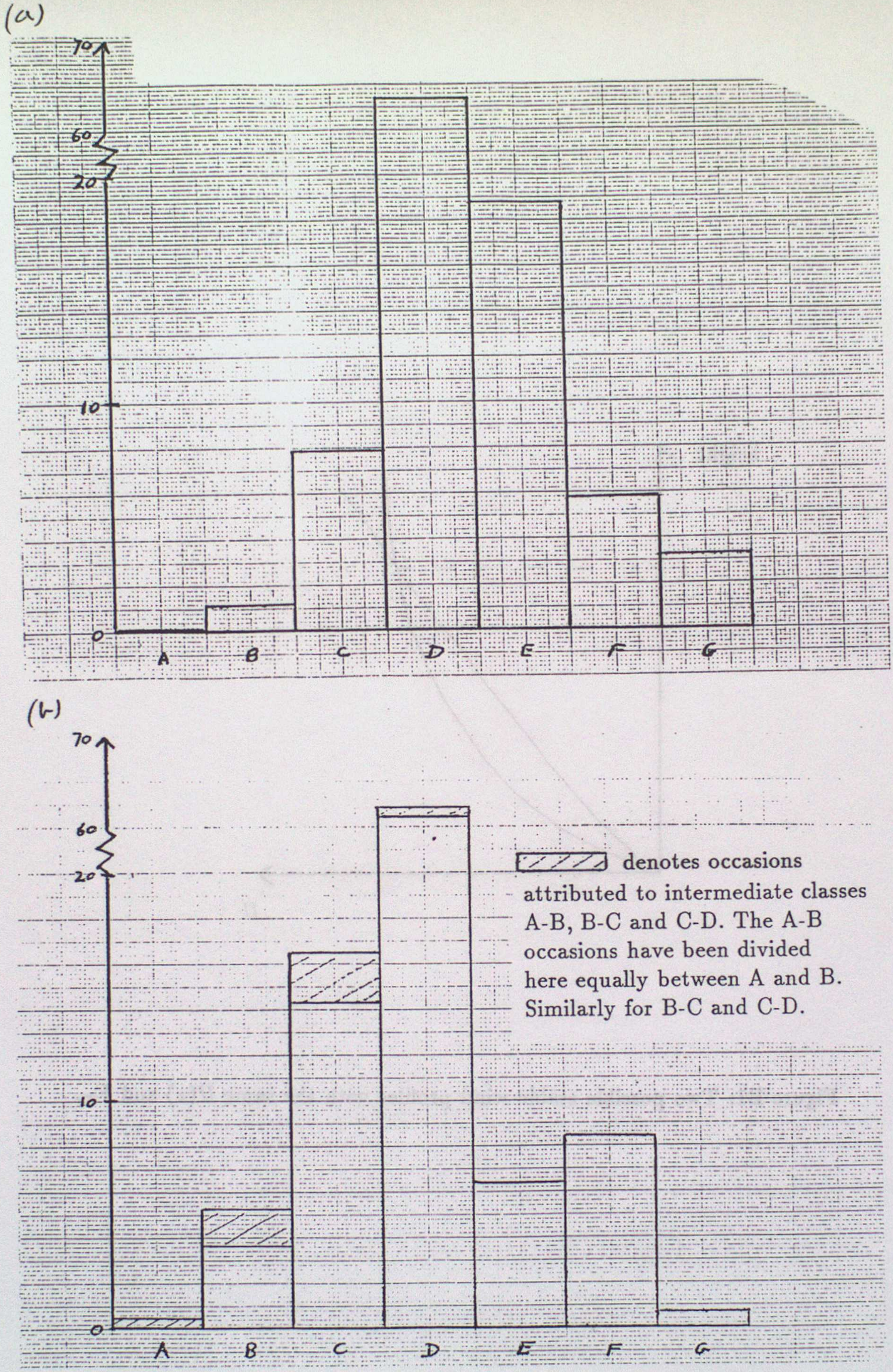


Figure A2: Frequency of Pasquill stability at Belfast (Aldergrove) calculated by the two methods described in the text. Figure A2(a) shows the results of the first scheme and figure A2(b) the results of the second scheme.

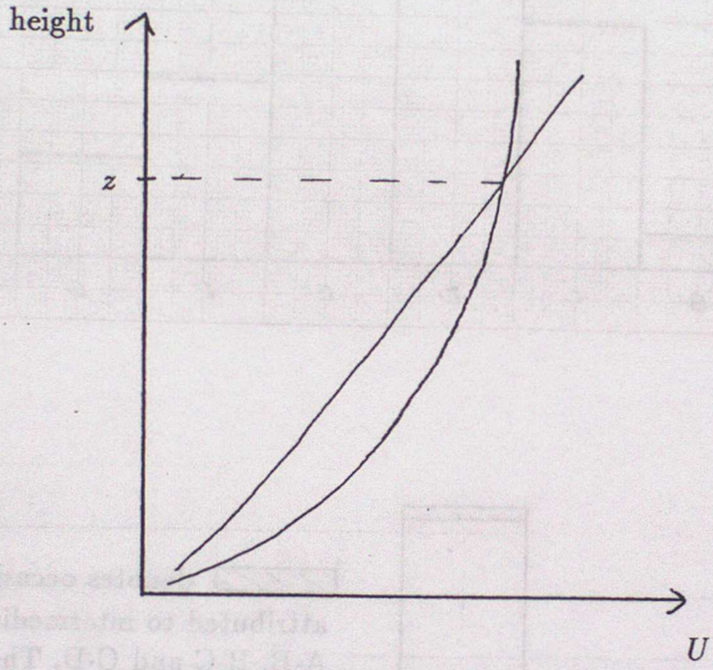


Figure B1: Two possible stable wind profiles with the same $U(z)$ and $F_{\theta 0}$.