CHAPTER-4

CONCEPTUAL DEVELOPMENT OF THE CTDM MODEL

In the previous chapter various modeling techniques were discussed. In this chapter, the conceptual development of the model is discussed.

4.1 Atmospheric Transport and Diffusion

Atmospheric transport and diffusion are controlled primarily by wind and stability. Transport is controlled by organized air motions or circulations which may be micro, meso or macroscale in extent. Diffusion is caused by turbulent eddies which are present due to surface heating (i.e. convective turbulence of vertical wind shear (i.e. mechanical turbulence).

The above description of atmospheric flow is adopted for computational and conceptual convenience and is not based on any natural flow scales. The division of the flow field into mean and turbulent components is arbitrary and is governed by the scale of averaging employed and by the nature of the phenomena being studied. On a sufficiently large scale, the mean flow field itself may be seen to possess its own turbulent fluctuations.

4.2 Scales of Motion

Three main scales of atmospheric motion are generally categorized: microscale, mesoscale and macroscale. The microscale includes atmospheric motion or phenomena with characteristic horizontal length scale upto about 2 km (Orlanski, I. et al., 1975 and
Atkinson, B.W., 1981). The corresponding time scales for the microscale range from fractions of a second up to an hour. Motion systems such as boundary-layer eddies in-cloud circulations, tornadoes, building wakes and thermal belong to the microscale. Other commonly used terms such as "local scale", "convective scale" small scale or "molecular scale" refer to at least a portion of the microscale.

The second main atmospheric scale, the mesoscale encompasses atmospheric motions or phenomena ranging from 2 to 2,000 km in horizontal extent. Corresponding time scales vary between one hour and one or two days. The mesoscale had been somewhat ignored by meteorologists historically due to a lack of instrument systems suitable for observing this scale. However, such important circulation systems such as sea or lake and land breezes, mountain and valley winds, low level jets and frontal circulations, squall lines are all mesoscale in extent. The terms "sub synoptic scale" and "regional scale" are sometimes used to refer to the upper end of the mesoscale (meso α), while the term "urban scale" is used in connection with smaller urban circulations (meso γ).

The third scale of atmospheric motion is the macroscale. It covers atmospheric circulation systems with characteristic horizontal length scales of 2,000 km or more. Corresponding time scales range from days to months. The most common atmospheric features on the macroscale in the mid-latitudes are the baroclinic eddies known as extratropical cyclones and anticyclones (or lows and highs). These systems are responsible for much of the day-to-day weather outside of the tropics and are often referred to as synoptic-scale. The largest atmospheric phenomena are the planetary waves. These immense flow structures may influence the weather of huge areas for months on end.
4.3 Microscale dispersion

In general, atmospheric dispersion on the microscale is dominated by turbulent diffusion and takes place within the confines of the planetary boundary layer.

Important factors influencing turbulent diffusion on the microscale include stability, the nature of the underlying terrain and local topography. Convective and mechanical turbulence contribute to diffusion on this scale in varying degrees.

4.3.1 Diffusion in the Neutral surface layer

The surface layer implies the lowest few tens of meters of the PBL. This layer typically has pronounced gradients of wind, temperature and moisture. The simplest surface layer is one which is horizontally homogeneous stationary and neutrally stratified. Horizontal homogeneity implies that the mean properties of the surface layers are independent of horizontal position and only depend on vertical distance $Z$ and time $t$. Ideally speaking, these surfaces should be infinitely uniform.

Stationarity implies that the properties of the surface layer are steady and unchanging with time. Such a state is unlikely to be exactly attained in the PBL due to the regular changes caused by the diurnal (i.e., 24-hour-or daily) cycle. However, if changes with time are so small as to be dynamically negligible, the surface layer may be said to be quasi-steady.

Neutral stratification implies a dry adiabatic lapse rate throughout the depth of the PBL. For this to occur, surface fluxes of heat and moisture must be negligible and no capping inversion should be present aloft. Given the diurnal cycle of surface heating and cooling conditions approximating neutral stability will likely occur only just after sunrise,
just before sunset or on windy overcast days. At such times, turbulence will be entirely due to vertical wind shear (i.e. to the change in wind velocity with height) and no buoyancy effects will be present.

4.4 Mean Wind Speed Profiles

The presence of a solid fixed boundary (the earth's surface) at the bottom of the surface layer slows the flow of air near the boundary. This effect is strongest close to the ground and becomes less and less important at heights farther and farther removed from the surface. As a result the mean transport wind speed \( U_* \) varies with height \( Z \). This variation under neutral stability conditions may be described by the formula:

\[
U(Z) = \frac{U_*}{k} \ln \frac{Z}{Z_o} \quad \text{valid for } z >> Z_o
\]

Where,

\( U_* \) is the friction velocity, \( k \) is Von Karman's constant and \( Z_o \) is the surface roughness or roughness length. The friction velocity is a measure of surface stress since it is defined as

\[ U_* = \left(\frac{\tau_o}{\rho}\right) \text{ where } \tau_o \text{ is the surface stress (drag) and } \rho = \text{ the ambient air density.} \]

In general \( Z_o = h/10 \) where \( h \) is the roughness element heights \( k = 0.4 \).

4.5 Diffusion in the Non-Neutral Surface Layer

Neutral conditions occur relatively infrequently in the surface layer. At other times stable or unstable conditions prevail. A stably stratified surface layer may occur at night
over land due to nocturnal cooling or at any time when warm air passes over a cold land or water surfaces. Turbulent eddies are damped in a stable surface layer, and the mean wind profile is a modified form of the logarithmic profile namely,

$$U(Z) = \frac{U_*}{k} \left[ \ln \left( \frac{Z}{Z_o} \right) - \psi \left( \frac{Z}{L} \right) \right]$$

Where the empirical integrated surface-layer similarity function $\psi$ is given by the formula

$$\psi \left( \frac{Z}{L} \right) = -4.7 \frac{Z}{L}$$

$L$ (Monin-Obukhov length) = $-\frac{U_*^3 \rho C_p T}{kgH}$

Where the symbols have their usual meanings. $L$ serves as a scaling length for diabatic situations (i.e., cases with a non-zero surface heat flux). The absolute value of $L$ may be thought of as the depth of the mechanically mixed layer near surface (Tennekes, H., et al., 1973). $L$ is small for very stable and very unstable conditions. The adoption of the non-dimensional height $\xi = Z/L$ as the fundamental parameter in diabatic surface layers is the basic premise of what is called Monin Obukhov similarity theory (Tennekes, H., 1982) or surface-layer similarity theory.

An unstably-stratified surface layer may occur during the day over land due to solar heating or at any time when cold air passes over a warm land or water surface. Under such conditions turbulent convective circulations occur and the convectively driven PBL may extend up as high as several kilometers. The mean wind profile in the unstable
surface layer is also a modified form of the logarithmic profile, but the empirical integrated surface layer similarity function has a different form namely (Benoit, R., 1977)

\[
\psi\left(\frac{Z}{L}\right) = -\ln\left(\frac{(\zeta^2 + 1)(\zeta_o + 1)^2}{(\zeta^2 + 1)(\zeta + 1)^2}\right) - 2[\tan^{-1}(\zeta) - \tan^{-1}(\zeta_o)]
\]

where

\[
\zeta = \left(1 - 15 \frac{Z}{L}\right)^{1/4}
\]

\[
\zeta_o = \left(1 - 15 \frac{Z_o}{L}\right)^{1/4}
\]

Thermal stratification has considerable influence over the boundary layer wind profile. Night time wind speeds tend to be smaller near the surface and greater higher up in the boundary layer; the reverse is true for daytime conditions. The daytime wind profile is also steeper near the surface than the corresponding night time profile resulting in greater wind shear and greater mechanical turbulence.

4.6 Turbulence parameters

The parameters \(\sigma_u, \sigma_v\) and \(\sigma_w\) are the standard deviations of turbulent velocity fluctuations in the x, y and z directions respectively. These quantities provide a measure of turbulence intensity. Formulas for \(\sigma_u, \sigma_v\) and \(\sigma_w\) for unstable neutral and stable conditions in the surface layer have been summarized by (Hanna et al., 1982). The formulae are based on surface layer similarity theory. Surface-layer turbulence scaling is discussed by (Tennekes, H., 1982) and much valuable material given by (Pasquill and Smith, 1983).
4.7 Diffusion in the Ekman Layer

The surface layer, although important for diffusion of pollutants emitted near the surface comprises only a fraction of the PBL. The upper part of the PBL is often called the Ekman layer or outer layer. This layer commonly makes up approximately 90% of the total PBL thickness (McBean et al., 1979). The Ekman layer differs from the surface layer in two respects: turbulent vertical fluxes vary with height and the mean vector "turns" with height. In contrast, within the surface layer, turbulent fluxes of momentum, heat and moisture remain nearly constant with height compared to their values at the surface (hence the name constant flux layer).

Coriolis effects are negligible within the surface layer and the mean wind direction tends to be constant with height due to mechanical and convective mixing. In the northern hemisphere, the mean wind vector traces out spiral curve as it turns to the right and just opposite occurs in the southern hemisphere. These changes in wind direction occur due to a changing balance between pressure gradient, friction and coriolis forces with height.

Variation of wind direction with height in the Ekman layer is another type of vertical shear (the first being changes in wind speed), and it results in increased lateral spread or dispersion of air pollutants. The angle between the "free wind" or "gradient wind" at the top of the PBL and the surface wind (usually measured at 10 meters) is sometimes called the cross-isobar angle. The cross-isobar angle increases with (1) increasing vertical stability (2) increasing surface roughness and (3) decreasing latitude. An average value of this angle for neutral conditions over average terrain at middle latitudes (45°) is 35°, but it can range from 10° (for unstable conditions, open ocean, 70° latitude) to 60° (for stable conditions, rugged terrain, 20° latitude) (Haltiner et al., 1957)
4.8 Effects of Stratification

The structure of the Ekman layer like the surface layer depends strongly upon stability. In a convective Ekman layer, surface heating causes random turbulent eddies but also drives recognizable discrete circulations such as heat plumes (or thermals), roll like horizontal vortices, and vertical vortices such as dust devils. Wind, specific humidity and potential temperature profiles tend to be very flat in the well-mixed convective Ekman layer, which is usually "capped" by an overlying layer of thermally stratified air. Turbulent motions are most intensive in the middle of the convective PBL because the earth’s surface inhibits motions near the base while the capping "inversion" layer acts to damp convective motions near the top.

The stably-stratified Ekman layer behaves quite a bit differently from the neutral or unstable Ekman layers. It is usually an order of magnitude shallower ranging from a few tens to several hundred meters in thickness. Turbulence levels are much lower since the larger eddies are strongly damped. The stable stratification also leads to propagation of internal gravity waves, a feature not seen in the other two types of Ekman layer. Because turbulence is damped in the stable PBL, it tends to respond slowly to changing conditions. As a result, the stable Ekman layer over land may be expected to be unsteady much of the time (McBean et al, 1979).

Similarity theory has also been developed for the Ekman layer, although for more restricted conditions. Our state of knowledge concerning the Ekman layer is considerably less advance than for the surface layer. One reason is the greater complexity of the Ekman layer.
4.9 The Mixing Layer

Mixing height defined as the near surface vertical height through which dispersion takes place is one very important parameter for the dilution of air pollutants. The mixing height is effectively equal to the PBL since most often the PBL is capped by a thermal inversion which inhibits further upward diffusion and mixing.

Within the mixing layer are generated characteristic profiles of temperature, wind speed, wind direction, shearing stress, heat flux, r.m.s. turbulent velocity, scale of turbulence. PBL height varies with vertical stability. A ground based convective or unstable "mixing layer" commonly occurs over land areas during the day and tends to reach a maximum depth in the early afternoon. At night the formation of a ground-based temperature inversions (i.e., the nocturnal inversion) over land areas corresponds to the transition to a shallower, stratified mixing (or mixed) layer. After sunrise the next day, solar heating initiates convection and the depth of the mixed layer increases again as convective motions erode the base of the capping inversion.

In a well-mixed boundary layer, pollution concentrations are inversely proportional to the depth of the layer. Hence, maximum dilution will occur during the afternoon as a rule and minimum dilution at night. As might be expected, maximum heights tend to be highest in the summer and lowest in the winter. The wind speed in the mixed layer also influences pollutant concentrations as the wind will carry a greater volume of air above the pollutant source. The product of the mean wind speed in the mixed layer and the mixing height can thus be interpreted as an overall ventilation or dilution factor (Holzworth, G.C., 1974) and is referred to as the ventilation coefficient and

\[ X \propto \frac{1}{UH} \]
4.10 The Mixing depth

The vertical extent of the PBL is quite variable. Its thickness can be characterized in a number of ways. The depth of boundary layer (h) is the thickness of the turbulent region next to the ground. This is also called the depth of the mixed layer, or the mixing depth since atmospheric properties are well mixed within it.

Another height used to describe the thickness of the PBL in the daytime or at night over heated surface is the height $Z_i$ of the lowest inversion. Typically, the temperature decreases with height in the daytime PBL and increases with height above. In an inversion (a region having temperature increase with height), turbulence is being suppressed. Thus, roughly $h$ and $Z_i$ are the same in the daytime. Actually, however $h$ tends to be 10% or so larger than $Z_i$ because the lowest part of the inversion is still turbulent, partly because of overshooting from below, partly because there is often a strong wind shear in the inversion.

At night an inversion often extends to the ground because the ground cools rapidly by emitting infrared radiation. When the wind is strong, mechanical turbulence is created and heat is lost to the ground by turbulent mixing throughout the boundary layer. However, on clear nights with weak winds, only the bottom of the boundary layer is turbulent. The upper part cools by divergence of flux of infrared radiation. Under such conditions the mixing depth $h$ (the turbulent region) and the boundary layer depth $Z_i$ (the top of the cooled region) may be very different from each other (Note that $Z_i$ now the top of the ground-based inversion.)
4.11 Neutral Conditions

The depth of the mixed layer in neutral stratification (Temperature decreasing upward at the isentropic rate), which occurs, for example, with very strong winds on cloudy days or nights, depends on the wind and terrain roughness.

Truly neutral PBLs are quite rare. According to rather simple arguments, \( h \), under such conditions should be proportional to the surface friction velocity \( U_* = \sqrt{(\tau / \rho)} \)

Where \( \tau \) is the surface stress and \( \rho \) air density, and inversely proportional to the Coriolis parameter, \( f \), defined by

\[
f = 2.0 \cdot \Omega \cdot \sin \phi \tag{1}
\]

Where, \( \Omega \) is the earth's rate of rotation \( (7.29 \times 10^{-5} \text{ sec}^{-1}) \) and \( \phi \) is the latitude.

The constant of proportionality is estimated to be 0.15-0.25 so that we have

\[
h = 0.25 \cdot U_*/f \tag{2}
\]

The reason for the effect of earth's rotation is that it influences the vertical wind shear and, therefore, the intensity of turbulence.

The friction velocity in neutral conditions is given by

\[
U_* = \frac{k_a V_d}{\ln \left( \frac{a}{Z_o} \right)}
\]

Where \( V_d \) is the wind speed at level \( a \) and \( Z_o \) the roughness length, \( K_a = \text{Von Karman constt}, \) usually taken nearly 0.4.

Relation of \( Z_o \) to various terrain types has been given in table (4.1).

In our case \( Z_o \) has been assumed to be 30 m.
<table>
<thead>
<tr>
<th>$z_0 (m)$</th>
<th>RELATION OF $z_0$ TO VARIOUS TERRAIN TYPES (ESDU, 1974)</th>
</tr>
</thead>
<tbody>
<tr>
<td>10^{-1}</td>
<td>Many hedges</td>
</tr>
<tr>
<td>10^{-2}</td>
<td>Few trees, summer time</td>
</tr>
<tr>
<td>10^{-3}</td>
<td>Off-sea wind in coastal areas</td>
</tr>
<tr>
<td>10^{-4}</td>
<td>Calm open sea</td>
</tr>
<tr>
<td>10^{-5}</td>
<td>Ice, mud flats</td>
</tr>
</tbody>
</table>

- **Centers of cities with very tall buildings**: Very hilly or mountainous areas
- **Centers of large towns, cities**: Forests
- **Centers of small towns, outskirts of towns**: Fairly level wooded country
- **Many trees, hedges, few buildings**
- **Farmland**
  - Long grass (≈ 60 cm), crops
- **Airports (runway area)**
- **Fairly level grass plains**
- **Natural snow surface (farmland)**
- **Desert (flat)**
  - Large expanses of water
- **Snow-covered flat or rolling ground**
4.12 Unstable Conditions

With positive heat flux at the ground (sunny conditions) and some wind, we will have both mechanical turbulence, and heat convection. But following energy budget of turbulence, the generation of mechanical turbulence decreases rapidly with increasing height, because it is proportional to the vertical wind shear. In contrast, the generation of heat convection varies slowly with height therefore; the top of the daytime boundary layer is determined by the parameters describing heat convection particularly vertical heat flux at the surface.

Following Carson,(1973); Deardroff,(1974), the equation is

\[ Z_t^2 = Z_a^2 + Z_u^2 \]

Where,

\[ Z_a = \frac{U_0}{4f} \] and \( Z_u \) is the extra contribution due to buoyancy forces

\[ Z_u = 2.3H + 1.25 \times 10^{-3}H^2 + 1.63 \times 10^{-5}H^3 \]

4.13 Stable Conditions

On clear nights with weak winds, only the lower portions of the inversion layer are continuously turbulent. This is the region in which the Richardson number is less than its critical value of 0.25. Also, this is the layer in which pollutants are mixed and the top of which can be detected by acoustic sounders. The height of this layer is denoted by h. Unfortunately, the top of the inversion layer is also sometimes taken as the top of the PBL. This may be much larger than h. Stable boundary layers are notoriously complex showing considerable variability in time and space. Gravity flows down slopes, however gradual, and the difficulty of estimating the surface heat flux are two obvious problems.
Radiational transfer of heat away from or to the surface and the rather fickle nature of the critical Richardson number which marks the transition between turbulent and non-turbulent states are two other serious problems. These difficulties must always be borne clearly in mind when interpreting the results.

\[ Z_s = 21500 \times U^2 \cdot (H)^{-1/2} \]

To provide for continuity into the stable conditions, we have

\[ \frac{1}{Z_{ST}^3} = \frac{1}{Z_n^3} + \frac{1}{Z_s^3} \]

Mahrt et al., (1979) recognize three layers within the part of the atmosphere cooled at night. At the bottom is the mixed layer of thickness \( h \). Next comes a region without turbulence or with some intermittent turbulence in which heat is transferred mostly by local wind shear. This three-layer structure has been modeled successfully by Andre et al.,(1978). Mahrt,(1983) has given a summary of recent papers in this area. Here, only the estimation of \( h \), the thickness of the lowest continuously turbulent layer has been treated. Qualitatively, we would expect \( h \) to depend on wind at a fixed height, roughness, stability and perhaps latitude.

There is some controversy as to whether simple equations for \( h \) as a function of these variables can be designed. Possibly only an equilibrium value \( h_e \) can be described with simple equations (e.g. Nieuwstadt, 1981). The actual value of \( h \) may have to be determined by a rate equation because mixing is slow.

\[ \frac{dh}{dt} = h_e - \frac{h}{\tau} \]
Where \( \tau \) is an appropriate time scale. Garrett (1982) suggests that the time scale, at least in middle latitudes is usually sufficiently small (strong enough mixing), so that \( h_e \) is a reasonable estimate of \( h \). The most common formula for \( h_e \) was first suggested by Zilitinkevich (Businger and Arya, 1974)

\[
h_e = a \sqrt{\frac{U_*}{f} L}
\]

Where \( a \) is a const of proportionality and \( L \) is the Monin Obukhov length.

According to the above equation, \( h_e \), approaches infinity in neutral air, becomes small in stable air, and depends directly on the square of \( U_* \) and inversely on the square root of heat flux \( H \) as

\[
L = - \frac{U_*^3 \rho C_p T}{k g H}
\]

\( h \) varies apparently as the square of the wind speed. Since, the friction velocity for a given wind speed increases with roughness, \( h \) increases with increasing roughness. Finally, the lapse rate affects \( h \) through the downward heat flux which is inversely proportional to \( L \).

Shortcomings of the Equation (1) Above

Attempts to test (1) by measurements, however, have been disappointing. There has been little correlation between measured values of \( h \) and those estimated by the equation (Mahrt et al., 1979); Garrett, (1982) suggest that the unsatisfactory results with equation (1) are due to difficulties in obtaining accurate measurements of \( U \) and \( L \) rather
than to inadequacies of the equation. He suggests that in middle latitudes the value of a in the above eqn (1) is about 0.4

Since estimates of \( U_* \) and \( L \) are unreliable during the relatively weak turbulence in stable stratification, it may be better for practical purposes to estimate \( h \) directly from measurements by acoustic radar (sodar). Sodar measures the distribution of high-frequency temperature fluctuations. Above \( h \), such fluctuations become intermittent; thus \( h \) can be inferred from the record. However, at night over land the mixing depth sometimes is so small that echoes from turbulence are masked by the ground clutter; further echoes may be received from the thin turbulent layers above \( Z_\i \) (inversion base). Thus, sometimes the height of the highest echoes is given as larger than \( Z_\i \) when, in fact, the true \( h \) is much lower. However, this ambiguity can be avoided by separating echoes from turbulent and laminar layers, which are different in appearance.

4.14 Atmospheric Boundary Layer

A boundary layer is defined as the layer of a fluid (liquid or gas) in the immediate vicinity of a material surface in which significant exchange of momentum, heat or mass takes place between that surface and the fluid. Sharp variations in the properties of the flow, such as velocity, temperature, and mass concentration also occur in the boundary layer. The atmospheric boundary layer is formed as a consequence of the interactions between the atmosphere and the underlying surface (land or water) over time scales of a few hours to about a day. Over longer periods, the earth-atmosphere interactions may span the whole depth of the troposphere, typically ten kilometers, although the PBL still plays an important part in these interactions. The influence of surface, heating etc. is
quickly and efficiently transmitted to the entire PBL through the mechanism of turbulent transfer downward through the PBL to the surface through the same mechanism.

The atmospheric PBL height varies over wide range (Several tens of meters to several kilometers) and depends on the rate of heating or cooling of the surface strength of winds, the roughness and topographical characteristics of the surface, large scale vertical motion, horizontal advections of heat and moisture and other factors. In the air pollution literature, the PBL height is commonly referred to as the mixing depth since it represents the depth of the layer through which pollutants released from the surface and in the PBL get eventually mixed.

Following sunrise on a clear day, the continuous heating of the surface by the sun and the resulting thermal mixing in the PBL causes the depth to increase steadily throughout the day and attain a maximum value of the order of 1 km (range = 0.2-5 km) in the late afternoon, latter in the evening and throughout the night on the other hand the radiative cooling of the ground surface results in the suppression or weakening of turbulent mixing and, consequently, in the shrinking of the PBL depth to a typical value of the order of only 100 m (range 20-500m).

4.15 Turbulent Transfer Processes

Turbulence is responsible for the efficient mixing and exchange of mass, heat, and momentum throughout the PBL. In particular, the surface layer turbulence is responsible for exchanging these properties between the earth’s surface and the atmosphere. Through the mass exchange process between the earth’s surface and the atmosphere, the radiation balance and the heat energy budget at or near the earth’s surface are also significantly
affected. More direct effects of turbulent transfer on the surface heat energy budget are through sensible and latent heat exchanges between the surface and the atmosphere.

Turbulent transfer of momentum between the earth and atmosphere is also important. It is essentially a one-way process in which the earth acts as a sink of atmospheric momentum (relative to earth). In other words, the earth's surface exerts frictional resistance to atmospheric motions and slows them down in the process.

An understanding of turbulence structure throughout the boundary layer is essential for the prediction of pollutant dispersion from ground-level and elevated source. Elevated source plumes are much more dependent on the turbulent structure of the whole boundary layer than are ground-level source plumes, especially in the region of high ground level concentration. The surface sensible heat flux $H$ is the main parameter determining the generation of turbulence by buoyancy forces over land. Fig. (4.1) gives the picture of the radiative process that maintain the heat balance of the earth atmosphere system.

Estimation of the surface sensible heat flux can be made by suitable vertical velocity and temperature fluctuation high-response probes kept at several meters above ground or less accurately by using measurements of the vertical gradients of mean temperature and wind. In climatological studies or in other situations where research instrumentation or detailed analysis are not practical, some reasonable estimate of $H$ can be made using fairly readily observed parameters along the following lines.

(Monteith, 1973) has given Heat flux as
Fig. 4.1. The radiative processes that maintain the heat balance of the earth-atmosphere system.
\[ H = \frac{r_a + r_{st} - r_i}{(1 + \frac{A}{\gamma})} R \]

Where \( r_a \) is the aerodynamic resistance and over grass with \( Z_o = 15m \), and is given by,

\[ r_a = \frac{212}{u} (Z_m) \]

Where \( U(2m) \) is the wind speed at 2 meters.

4.14 Aerodynamic Resistance

Aerodynamic resistance is the descriptor of the external factors that can affect a plant's transpiration and is roughly related to the amount of time it takes for the vegetation to exchange water vapour with the air. It depends on the wind speed and the geometry and roughness of the plant community. Generally, as surface roughness increases, such as with the height of the vegetation, the turbulence becomes greater and the aerodynamic resistance becomes reduced. Similarly, increases in foliage density can cause resistance to increase. Assuming that increases in wind speed do not alter the shape and roughness of the plant community, aerodynamic resistance will be inversely proportional to the wind speed.

For near neutral (dry adiabatic) conditions, the aerodynamic resistance, \( r_a \), can be computed from the wind profile (Federer, 1979; Monteith, 1965; Oke, 1978; Van Bavel and Hillel, 1976)

\[ r_a = \frac{\ln \left( \frac{Z-d}{Z_o} \right)}{K^2 U} \]
Where \( r_y \) is in sec/cm, \( u \) is the wind speed measured at height \( Z \); \( d \) is zero displacement, determined experimentally, that allows a straight relationship between \( \ln(Z-d) \) and \( u \); \( Z_0 \) is the roughness parameter, which is equal to \( Z-d \) for \( u=0 \); and \( k=0.41 \) is the Von karmann constant. When the plane height is relatively uniform, the parameter \( d \) can usually be approximated as two thirds the height of the plant community.

4.15 Climatological Resistance \((r_i)\)

Climatological resistance is given by

\[
r_i = \frac{\rho C_p}{\gamma R} \left( 1 - \frac{h}{100} \right) e_w(T)
\]

Where, \( r_i \) is the so-called climatological resistance, \( h \) is the relative humidity of the air, \( e_w(T) \) is the saturated vapour pressure at temp \( T \), \( \rho C_p \) and \( \gamma \), the Psychrometric constant are also functions of \( T \) (Smith and Blackll, 1979), see Table (4.2). \( r_s \) is the stomatal resistance. \( \bar{\Delta} = \frac{\partial}{\partial T} e_w(T) \) is taken as the average between \( \Delta \) at \( T \) and \( \Delta \) at soil temp, \( T_s \).

Where,

\[
T_s = T + \left[ 5.25 \times 10^{-3} r_s - 6.25 \times 10^{-6} r_s^2 \right] \frac{R}{\rho C_p} r_s
\]

If \( T>25 \), \( r_s = 3000 \)

If \( T \leq 25 \), \( r_s = 2000 \)

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Table 4.1
List of physical costant$\$s

<table>
<thead>
<tr>
<th>T(°C)</th>
<th>$e_w(T)$</th>
<th>$\Delta$</th>
<th>$\rho$</th>
<th>$\rho C_p$</th>
<th>$\gamma$</th>
<th>L(J/g)</th>
</tr>
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<tbody>
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<td>-5</td>
<td>4.21</td>
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<td>1.316</td>
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<td>2513</td>
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<td>1.315</td>
<td>1315</td>
<td>0.645</td>
<td></td>
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<td>-1</td>
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<td>0.45</td>
<td>1.292</td>
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<td>0.646</td>
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<td>0.48</td>
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<td>1301</td>
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4.16 Canopy Resistance

Canopy resistance is a bulk physiological descriptor of the effects that the internal soil-plant system has on evapotranspiration rate. The canopy resistance, $r_c$, is most closely related to the stomatal resistance.

4.17 Stomatal Resistance

Stomatal resistance, also called the diffusion resistance, depends on both the physiological limitations of the stomata to the exchange water and the rate of flow through the soil-root-stem-leaf pathway, which is sometimes controlled by the availability of soil moisture.

When ample moisture exists in the soil to supply the plant with water, stomatal resistance depends mainly on the behaviour of the stomata in relation to the photosynthetic process. As long as the water supply is adequate, the stomata will open fully in response to sunlight to a level of minimum resistance. The stomata will close when the intensity of sunlight is low enough to limit photosynthesis (thereby increasing the stomatal resistance).

When the transpiration from the leaves exceeds the capability of the soil-root system to supply water to the leaves; the stomata begins to close, creating greater resistance to the transpiration process. The point at which this stress due to the soil moisture depletion occurs depends on the transpiration rate and the hydraulic properties of the soil-plant system. Often when a high radiative load is placed on a plant, stress may occur even at high soil moisture content because of hydraulic limitations of the plant and soil. Once this stage is reached, the transpiration rate is limited by the amount of water that
can be taken up by the roots.

The data measurement and analysis required for estimating $r_n$ are somewhat tedious. Diffusion parameters are available that measure the rate of water-vapour diffusion from a leaf, but numerous measurements should be made over the entire canopy, since resistance can vary even between layers of an individual plant (Brun et al., 1973; Morrow and Slatyer, 1971). Factors other than canopy resistance also affect stomatal behaviour, the most notable of which is the density of plant cover, usually represented by the leaf area index (Brun et al., 1973; Black et al., 1970). A common technique for estimating canopy resistance is to divide the stomatal resistance by the leaf area index, though it is usually agreed that the relationship among the three parameters is more complex (Federer, 1979). The common techniques for estimating $r_e$ are not exact, and it is fortunate that when aerodynamic resistance is nearly equal to or greater than canopy resistance, the estimated evaporation is relatively insensitive to error in estimation of $r_e$.

4.18 Determination of Net Radiation ($R_n$) and Soil Heat Flux ($G$) (Smith, F. B. and Blackall, R. M., 1979)

$R$ is the heat available at the ground surface to go into the atmosphere and is equal to the difference between the net incoming radiation $R_n$ and the soil heat flux $G$. $R_n$ depends mainly on solar elevation and cloud cover. Solar elevation is given by $\theta$ where

$$S = \sin \theta = \sin \phi \sin \delta + \cos \phi \cos \delta \cos \frac{\pi}{12} (t - 12)$$

Where $t =$ time in hours (Local Apparent Time)
Local Apparent Time for a given standard time is written as

Local Apparent Time = Local Mean Time + Equation of Time

= Local Standard Time + Longitudinal Correction + Equation of Time

= Local Standard Time + 4(Ls - Es)

Where Ls is the Standard Longitude

and \( \delta \),

The Solar declination is given by,

\[
\delta = 23.3 \sin \frac{2\pi(n-81)}{N}
\]

Where,

\( n \) = Julian day number and \( N \) = no. of days in a year, \( \phi \) is the latitude.

The analysis of values of \( R \) at Cardington during 1976 suggest that \( R_a \) has surprisingly little variation with cloud as long as the amount is \( 6/8^{th} \) or less. Clearly reflected light off cloud, diffuse radiation and long wave radiation from cloud tends to compensate for the reduction in the direct solar radiation as cloud increases. However, for \( 7/8^{th} \) cloud or completely overcast conditions, \( R_a \) significantly drops. In the former case

\[
R_a = 925S^2 - 229S^3 - 17 - 90.8S
\]

While in the latter

\[
R_a = 24 + 68.35S + 50S^3
\]

Actual values show some scatter about these lines. The standard error being 30W/m². Over the sea or in coastal areas the clearer atmosphere would be expected to give
values some 50% higher than predicted above.

The soil heat flux $G$ is not readily measured requiring sometimes dubious extrapolation from measurement using heat a flux plates at different levels within the soil. Cardingtion measurement is taken at face value imply the following impressions valid upto 1 hour before sun set.

$$G = R_a(1.6\sin^{1/2}x + 0.292\sin^{1/4}x - 0.229\sin x)A$$

Where, $X = \pi(4t-3t_o-12)(152-14t_o)^{-1}$

$t_o$ = time in hours of sunrise

$A = 0.2$ if sky is not overcast

$= -0.08$ if sky is overcast.

Although this determination of $H$ seems rather complicated, once the individual steps are understood, it is very straightforward. Experience shows that a correlation between forecast and actual values of $H$ is as high as 0.9 and the standard error is only about 36 Wm$^{-2}$. Surface roughness $Z_o$ characterizes the dynamic roughness of the ground and is related to the physical nature of the roughness elements. Table gives some typical values.

4.19 Diffusion Coefficients

4.19.1 The vertical cloud standard deviation ($\sigma_z$)

Vertical diffusion of pollutant within the atmosphere is a consequence of turbulent motions or eddies superimposed on the mean airflow. The more intense these eddies are and the larger they are, the more rapid is the dispersion. Ground level concentration at first increases by rapid dispersion, if the source is elevated but later decreases. For
ground level sources, the surface concentrations are always reduced by rapid vertical dispersion.

There are two normal sources of turbulent energy. The first is dynamic in nature and arises from the breaking action of a rough underlying surface on the airflow. This results in momentum and energy being transferred from the mean motion into the tumbling eddies. The eddies in turn help to bring down mean motion momentum from aloft to balance the losses in the surface layers. The intensity of turbulence increases both with wind speed and with roughness $Z_o$ of the underlying surface. One measure of the intensity is the friction velocity $U_*$ defined as:

$$U_* = \sqrt{\tau / \rho}$$

where $\tau$ is the surface drag/unit surface area; $\rho$ is the density of the air.

In neutral stability conditions, the intensity of the vertical motions, represented by the root mean square vertical velocity $\sigma_w$ is directly related to $U_*$ in the following manner $\sigma_w = 1.3 U_*$ generally.

The second major source (or sink) of turbulent energy is the buoyancy generated by internal density or temperature differences. The differences arise from the air and the underlying ground surface having different temperatures and water vapour pressures. Unlike the situation over the sea, the buoyancy effects arising from water vapour over land can, normally, be neglected and attention can be focussed on the consequential flux of sensible heat either from the ground into the air (when the former is hotter than the latter in the day) or the reverse (as often happens at night).

During the day incoming solar radiation tends to heat the ground and some fraction of this may be advected into the overlaying airflow as sensible heat. The
elevation of the sun, the amount and type of cloud and the dampness and character of
ground surface are obvious factors determining the upward heat flux H. At night, the
ground usually cools as a result of imbalance between outgoing and incoming long wave
radiation. The incoming radiation component is largely governed by the amount of cloud,
and is, therefore, probably the principal variable determining thermal influence of the
ground on the air at the night.

When the sensible heat flux H is upwards (from ground to the air), temperature
tends to increase rapidly in a downward direction and, in consequence, any fluid element
perturbed upwards, say, soon finds itself hotter than its environment and buoyancy
accelerates it upwards. The motion is unstable, turbulent motions tend to be intense, and
pollution dispersion is rapid. At night when the heat flux is downwards, the temperature
decreases downwards and perturbed fluid elements are soon restored back to their original
levels. Turbulence may be entirely quenched in the night, especially under clear sky
conditions when surface cooling is rapid and in light winds when the dynamic generation
is small.

At a given distance from the source, \( \sigma_z \) is a function of the atmospheric stability
donw wind distance and ground roughness. The mathematical description is as follows:

\[
\sigma_z = \frac{x u^* b}{U(Z_m)}
\]

Where

\[
U_Z = \frac{U^*}{\kappa} \ln \frac{Z}{Z_o}
\]

\[
Z_m = Z_s + \Delta h \quad \text{where symbols have their usual meanings.}
\]

\[
b = 1.3 \left( 1 - 0.8 \frac{Z_m}{h} \right) \left( 1 + \frac{t}{2T_L} \right)^{-1/2}
\]

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\[ t = \frac{x}{U(Z_m)} \quad , \quad T_L = \frac{\Lambda_x^{(w)}(Z_m)}{\sigma_w(Z_m)} \]

\[ \sigma_w = 1.3U_\star (1 - 0.8 \frac{Z_m}{h}) \]

\[ \frac{1}{\Lambda_x^{(w)}} = \frac{2}{Z_m} + \frac{3}{h} \]

Where,

\( \sigma_z \) is the root-mean-square vertical velocity which represents the intensity of vertical motion in neutral stability conditions.

In the derivation of \( \sigma_z \), the vertical diffusion coefficient, one modification was introduced. When \( Z_m > h \) (i.e., effective stack height exceeds the PBL depth), we take \( Z_m = h \). This gives the maximum possible ground level concentration. The same was done in the determination of \( \sigma_y \). Since, when \( Z_m > h \), the formula used for their determination fails to calculate the values.

### 4.19.2 The Determination of the Crosswind Spread \( \sigma_y \)

The dispersion of the plume in the horizontal plane is the result of turbulence process together with fluctuations in wind direction. \( \sigma_y \) is dependent on horizontal turbulent fluctuations over various time scales (i.e., on microscale, masoscale and large scale eddies and on the interaction between the vertical turbulent fluctuations and shear in the mean horizontal wind velocity. In the near field, when a plume has not spread deeply through the atmospheric boundary layer, this last term is not important and it will
not be considered explicitly in the following. However, it may be the reason that
measurements of $\sigma_y$ against distance indicate a distance dependence which is not a power
law of distance with a simple exponent. The way in which $\sigma_y$ depends on eddy sizes and
on averaging time and travel time can be seen from Tailor's theory of diffusion in
stationary homogeneous conditions.

$$\sigma_y^2 = 2\sigma^2 \int_0^1 \int_0^t t R(t_2) dt_2 dt_1$$  \hspace{1cm} (1)$$

Where,

$\sigma_y$ is the standard deviation of horizontal wind fluctuations and $R(t)$ is the
Lagrangian auto-correlation of the horizontal wind velocity fluctuations. $R(t) = 1$ for
small $t \ (t/t_L \leq 1)$ and zero for large $t \ (t/t_L > 1)$, where the Lagrangian time scale, $t_L$, is
defined by

$$\int_0^\infty R(t)dt$$

The usual assumption that is made is that a major contribution to the crosswind
spread is from eddies with a Lagrangian time scale long compared with travel time, so
that

$$\sigma_y^2 = \sigma_y^2 t^2 b^2 + \sigma_{y_t}^2$$  \hspace{1cm} (2)$$

Where $\sigma_{y_t}$ is the turbulence-induced term and $b^2$ is an empirical correction
hopefully close to unity or alternatively,

$$\sigma_y^2 = \sigma_{y(t)}^2 + \sigma_{y(\infty)}^2$$
where,

\[ \sigma_y(t) = \frac{2xu_z}{U(Z_m)} \left( 1 - 0.8 \frac{Z_m}{h} \right) \]

\[ \sigma_y(w) = 0.065x \left( \frac{7T}{U(Z_m)} \right)^{1/2} \]

4.20 Plume Behaviour in the Atmosphere

As pollutants exit from a stack and enter the ambient air, they are influenced by many variables, such as exit temperature, momentum and composition of the effluent gases, and the horizontal momentum, temperature, and turbulence of the ambient air. As the ambient wind speed increases across the exit of the stack, effluent gasses become more dilute. Superimposed on this dilution is diffusion due to turbulent eddies in the ambient air.

A plume dispersing in a field of small turbulent eddies moves in a relatively straight line with a gradual increase in Plume cross-section. If a plume disperses in a field of large eddies (large compared to the diameter of the plume), it grows very little in size but meanders widely. The typical daytime atmosphere contains eddies of an infinite variety. The plume pattern varies with the thermal structure of the atmosphere.

4.21 Plume Rise

The formula used for predicting the rise of buoyant plumes was developed by D.J. Moore (1980). The distinct aspect of this formula is that it recognizes that the plumes
often break up during the plume rise phase into blobs that are only rather loosely interconnected. This implies three dimensional mixing with the environment rather than the two dimensional mixing implied implicitly by Briggs. Consequently, the rise is proportional in Moore's formula to $Q_{14}$ rather than $Q_{13}$. Moore's formula is rather more empirical than Briggs and attempts are being made to interpolate between rise conditions in various meteorological situations.

Mathematical description of the formula is as follows

1. $Q_h$ is the heat flux in mega watts in the stack gases. $Q_h$ may be calculated directly from the efflux velocity $V_s$ of the gases at stack top, the stack top diameter $d$ and temperature difference $(T_s - T_o)$ between the stack gases and the environment.

$$Q_h = (\frac{\pi}{2}) \rho C_p d^2 V_s (T_s - T_o) \times 10^{-6}$$

2. $U$ is the wind speed ($m/s$) at stack height, $g = 9.81$ (m/s$^2$) is the acceleration due to gravity, $h_s$ is the stack height in meters and $\Delta \theta$ is the increase in potential temperature of the environment of air over 100 m above the stack top. Then let

$$U^* = \max(0.2, U)$$

$$\Delta \theta^* = \max(\Delta \theta, 120)$$

$$h^* = \min(h_s, 120)$$

$$f = 1 \text{ if } \frac{\Delta \theta}{U^2} > 0.0025$$

$$0.16 + 0.007 h^* \text{ otherwise}$$

$$x_N = \min(4224, 1920 + 19.2 h_s)$$

$$T' = \max(120, T_s - T_o) = \theta_0^*$$

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The plume rise $\Delta h$ is as a function of downwind $x$ is then given by

$$\Delta h = 2.25 Q^{1/4} \cdot x^{3/4} \cdot \frac{F}{U^*} \left( \frac{T'}{110} \right)^{1/8} \left[ 1 + \frac{27d + 1.5x_m}{x^*} + \frac{54dx_m}{x^*} \right]^{1/4}$$

or in another form

$$\Delta h = 0.922 \frac{f}{u_1} \left( \frac{\theta_o^*}{110} \right)^{1/8} \left[ \frac{Q_h}{C_p \theta_e} x^* (x^* + 27d_o) \right]^{1/4}$$

where $\Delta \theta^* = \gamma$

$\gamma_{\text{stable}} = 1.25^\circ \text{k}/100\text{m}$

$\gamma_{\text{unstable}} = 8^\circ \text{k}/1000\text{m}$

$\gamma_{\text{neutral}} = 5^\circ \text{k}/1000\text{m}$

$\theta_e = \text{(potential) temperature of air outside the plume at the stack height.}$

4.22 Terrain Adjustment

Consideration of the differences between ground level elevation at the stack and the receptor is an option in the complex terrain diffusion model. Terrain adjustment factors are included for these stability classes, i.e., stable, neutral and unstable. These factors can be any number between 0 and 1, calculations are made by subtracting the difference, defined as the elevation of receptor ground level minus the elevation of the
source ground level, from the effective plume height.

Inclusion of 1 as the adjustment factor means full response of the plume to the terrain factors, i.e., they simulate the plume rising over terrain features (the entry of 1 for the terrain adjustment for a given stability will compute as if the source and receptor were on a level surface. The equation used for the inclusion of the terrain adjustment is

\[ H_A = H - \Delta E + F_T \Delta E \]

where,

- \( H_A \) = adjusted effective height
- \( H \) = effective height
- \( \Delta E = E_R - E_S \)
- \( E_R \) = ground level elevation of receptor.
- \( E_S \) = ground level elevation of source.
- \( F_T \) = Terrain adjustment factor.

If \( F_T \) is put 0, it means no effect of the terrain features on the plume behaviour. This option is used for stable atmospheric conditions. And \( F_T = 0.1 \) for neutral atmospheric stability. Fig. (4.2).

4.23 Consideration for the Variation of Wind Speed. (Irwin, 1974)

In this model the input wind speed is assumed representative for the input anemometer height above the ground. The wind speed \( U_h \) at the physical stack height is calculated from:

\[ U_h = U_z (h/Z_a)^p \]
Fig. 4.2  Adjustment of plume height due to terrain.
where $U_z$ is the wind speed for the hour, $Z_a$ is anemometer height and the exponent $p$ is a function of stability.

For stable case, \( p = 0.65 \cdot 5.5 \)

for neutral case, \( p = 1.5 \)

for unstable case, \( p = 0.67 \cdot 1 \)

4.24 Concentration Computations

Concentrations are calculated as follows:

\[
\frac{C}{Q} = \left( \frac{1}{2\pi \sigma_y \sigma_z U} \right) \exp \left( -\frac{y^2}{2\sigma_y^2} \right) \left\{ \exp \left( -\frac{(Z-h)^2}{2\sigma_z^2} \right) + \exp \left( -\frac{(Z+h)^2}{2\sigma_z^2} \right) \right\}
\]

where $C$ is the concentration at the receptor in $\mu g/m^3$; $Q$, the pollutant load (emission rate), ($\mu g/sec$). $U$ is the wind speed and $\sigma_y$ and $\sigma_z$ are diffusion coefficients in $y$ and $z$ directions in meters. $h$ is the effective height of source emission in meters and $z$ is the receptor height. \( h = h_A \)

In the next chapter the results are discussed followed by the final chapter which deals with the shortcomings of the model and suggestions for its improvement.