CHAPTER 3

MAJOR FACTORS AFFECTING AIR POLLUTION DISPERSION

3.1 Introduction

All air pollutants emitted by point and distributed sources are transported, dispersed, or concentrated according to meteorological and topographical conditions. The airborne cycle is initiated with the emission of the pollutants, followed by their transport and diffusion through the atmosphere. The cycle is completed when the pollutants are deposited on vegetation, livestock, soil and water surfaces, and other objects, when they are washed out of the atmosphere by rain, or when they escape into space. In some cases the pollutants may be reinserted into the atmosphere by the action of wind.

In those regions where the topographic and meteorological conditions are conducive to the accumulation and concentration of pollutants, as in the case of the Los Angeles basin, the pollutants may hasten the deterioration of buildings and adversely affect public health as well as vegetation in the area. During the time period time the pollutants are airborne they may undergo physical and chemical changes. Smog, with the associated eye irritation, is the result of the interaction in the atmosphere of the oxides of nitrogen, selected hydrocarbons, and solar energy. The results of such transformations are not always harmful, however; sometimes they are beneficial, as in the case of some mineral salts that are necessary for plant life.

In large urban areas pollutants emitted from numerous concentrated sources, as well as distributed sources, are dispersed over the entire geographical area. Any given location within the urban area receives pollutants from the different sources in varying amounts, depending upon prevailing winds, presence of tall buildings, and so on. If the allowable concentration of a selected pollutant at a given location is not to be exceeded, the contributions made by different individual sources must be established.

The dispersion of a pollutant in the atmosphere is the result of three dominant mechanisms: (1) the general mean air motion that transports the pollutant downwind, (2) the turbulent velocity fluctuations that disperse the pollutant downwind, (3) mass diffusion due to concentration gradients. In addition, the general aerodynamic characteristics, such as size, shape, and weight, affect the rate at which the particles of nongaseous pollutants settle to the ground or are buoyed upward. The factors affecting both the wind and atmospheric turbulence will be discussed in the following sections of this chapter.

3.2 Global effects

3.2.1 **Solar Radiation**

The solar radiation is mainly a global scale factor affecting the atmospheric stability condition in an interested area. It is one of the parameters for determination of the atmospheric stability other than vertical temperature gradient, wind velocity, and cloudiness, etc. At the upper boundary of the earth's atmosphere the vertical solar radiation, termed the "solar constant," is approximately 8.16J/cm^{2}min. The maximum intensity occurs at wavelengths between 0.4 and 0.8 mm, which essentially is the visible portion of the electromagnetic spectrum. Approximately 42 percent of this energy is either (1) absorbed by the high atmosphere, (2) reflected to space by clouds, (3) back-scattered by the atmosphere, (4) reflected by the earth's surface, or (5) absorbed by water vapor and clouds. Approximately 47 percent of the solar radiation is absorbed by the earth's water and land surfaces. The earth, approximated as a body at roughly 290°K (62°F), radiates long-wavelength radiation with the maximum intensity between 4 and 12 mm (near-infrared region).

Insolation, or the quantity of solar radiation reaching a unit area of the earth's surface, is a function of many variables. The most important factor is the variation in the angle of incidence, as illustrated in Figure3.1. In Figure 3.1(a) the increase in surface area receiving the same quantity of insolation in the winter compared with the summer is illustrated. A similar change is illustrated in Figure 3.1(b) for different geographical locations

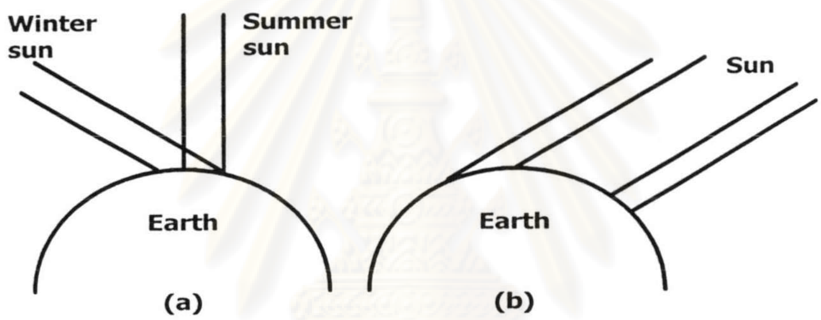

Figure 3.1 (a) Variation in insolation with different seasons. (b) Variation in insolation with geographic location. (Kenneth Walk, 1981)

The thickness of the atmosphere and thus the quantity of solar energy absorbed is a function of the time of day, as illustrated in Figure3.2. The sun's rays are tangent to the earth's surface in the morning and evening and approximately perpendicular at noon. The period of insolation for a summer day is approximately twice as long as for a winter day. From the preceding discussion we can see that the actual quantity of solar energy received by a unit surface area on the earth's surface is a complex function of location, season, time of day, and composition of the atmosphere above the surface.

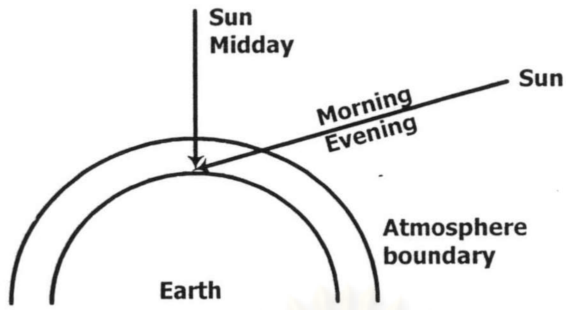

Figure 3.2 Variation in atmospheric thickness traversed by solar radiation with angle of incidence. (Kenneth Walk, 1981)

The specific heats of solid earth surface materials are lower than the specific heat of water. Thus, even though the quantity of solar energy absorbed by a unit surface area of land may be the same as that for a unit surface area of water, the resulting temperature increase will be different. Currents in water and the resulting heat transfer by convection cause the energy received at the surface to be transported to a greater depth in water than is the case in rock and soil, where heat transfer by conduction alone occurs. The combination of all the aforementioned effects produces the striking differences between water and land temperatures and hence between marine and continental air temperatures.

The wind patterns near the shore of bodies of water such as lakes, oceans, and bays are complicated by a factor nonexistent over land areas. Differences in the rate of warming between the land and water result in the development by midmorning of a distinct temperature difference between the air above the water and that above the land. The expansion of the rising warmer air over the land causes a general air movement horizontally from the water to the land (sea or lake breeze). At night the land surface cools at a faster rate by radiation than does the water. The air over the land gradually becomes cooler and denser than the air over the water. Hence the general horizontal air movement is from the land to the water (land breeze). Since major large cities are located near large bodies of water, the wind patterns are quite complicated, especially if mountains or large hills are nearby.

3.2.2 Wind Circulation

The general wind circulation mainly affects the wind direction. The sun, the earth, and the earth's atmosphere form one very large dynamic system. The differential heating of the air gives rise to horizontal pressure gradients, which in turn lead to horizontal motion in the atmosphere. Thus the temperature difference between the atmosphere at the poles and at the equator, and between the atmosphere over the continents and over the oceans, causes large-scale motions of the air. (Local winds such as lake breezes are caused by local temperature differences.) If the earth were not rotating, air normally would tend to flow directly from high-pressure regions toward regions of low pressure which in the horizontal usually means from a cold area toward a warm area. In Figure 3.3(a) the flow is shown to be perpendicular to the isobars. The rotation of the earth alters this situation. In addition of the pressure gradient F_p, one must also consider the Coriolis force F_{Cor} arising from the earth's rotation. (The Coriolis force is sometimes termed the horizontal deflection force.) This force accounts for the apparent deflection of a moving air parcel to the right in the Northern Hemisphere, relative to the surface, when the observer faces in the direction of motion of the parcel. The Coriolis force in this global system is a function of the velocity of the air parcel, as well as the latitude and the earth's rotational angular speed. It is a maximum at the poles of the earth and zero at the equator. If this vector force is added to the pressure gradient force, then in general the situation is represented by Figure 3.3(b), with the resultant velocity vector at some angle to the isobars. This representation is not one of static equilibrium, however, since the forces are not in balance as drawn. In the upper atmosphere, moreover, air parcels frequently experience relatively little acceleration. Therefore the forces acting on the parcel in this case must essentially be in balance. If we consider only the pressure gradient force and the Coriolis force to be present in the idealized case, the vector representation must appear as in Figure 3.3(c) for parallel isobars. Since the pressure gradient force F_p must be perpendicular to the isobars, the Coriolis force must be parallel to F_p but in the opposite direction, toward the high-pressure region. In addition, the wind velocity and the Coriolis force act at right angles to each other. As a result, the wind must blow parallel to the isobars. Moreover, recall that F_{Cor} acts to the right of the wind velocity in the Northern Hemisphere. Hence the wind must blow so that the low-pressure region is to the left of the direction of motion, as the observer looks downward toward the earth's surface. This idealized or conceptual wind is called the *geostrophic* wind by meteorologists. It's symbolized by V_g in the figure, and it approximates conditions a few hundred meters or more above the surface of the earth. Except in the case of very light winds, the direction and speed of the actual wind probably do not differ by more than 10 degrees and 20 percent, respectively, from the geostrophic values. For geostrophic flow the isobars coincide with the streamlines of flow.

Another type of wind referred to in meteorology is the gradient wind, which is associated with curved isobars. Even though the speed of an air parcel may still be constant, in a curved path the centripetal acceleration a_c must be taken into account. The geostrophic and gradient winds are concepts of practical interest in the absence of a significant frictional force. However, the movement of air near the earth's surface is retarded by frictional effects of the surface roughness. The vertical region between the earth's surface and the upper levels of the atmosphere where the gradient wind concept is valid is called the *planetary boundary layer*. The magnitude of the retardation of wind speed with height and the thickness of the boundary layer are functions of the surface roughness or terrain, as well as of the temperature gradient in the lower atmosphere. The effect of this frictional force, when added to the pressure and Coriolis forces, is to turn the air movement slightly to the left of the gradient wind (when the observer looks downward toward the earth's surface). In effect, the wind is turned at a slight angle toward the low-pressure region. The angular shift is a function of the same variables noted above for the frictional force. The vector diagram in Figure 3-4 illustrates the direction of the resultant wind for straight isobars. The frictional force is proportional to the wind speed. The frictional force directly reduces the wind speed in the boundary layer, thus reducing the Coriolis force, F_{Cor}. The pressure force F_p remains the same, however, so that it no longer is balanced by the Coriolis force as in a geostrophic wind. The result of this imbalance is that the direction of the wind is now across the isobars toward the low-pressure region, rather than parallel to the isobars. Note that the wind speed is now less than the geostrophic wind speed, all other conditions remaining the same.

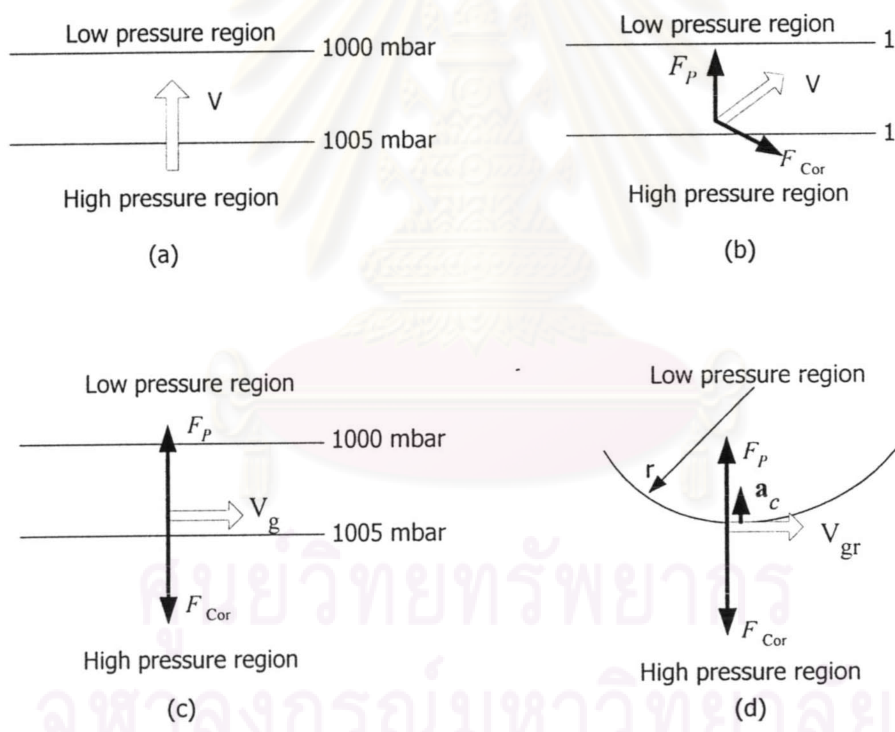

Figure 3.3 Effect of various forces on wind direction, relative to isobars in the atmosphere. (a) Pressure gradient force only, parallel isobars. (b) Pressure gradient and Coriolis forces, parallel isobars. (c) Pressure and Coriolis forces in balance, parallel isobars. (d) Pressure and Coriolis forces balanced by centripetal acceleration, curved isobar. (Kenneth Walk, 1981)

A steady wind can still exist when the frictional force is included in the analysis. The component of the F_p vector in the wind direction just balances the frictional force, F_f . At the same time, the reduced Coriolis force is just balanced by the remaining component of F_p. An interesting consequence of this is with respect to flow around high- and low-pressure centers, with friction. Figure 3-5 (a) shows the vector diagram for circulatory flow close to a high-pressure center. The pressure gradient, Coriolis, and frictional forces are shown, as well as the centripetal acceleration vector. This is merely a superposition of the diagram in Figure 3.4 upon a curved path. The important thing to note is that the wind velocity is outward from a circle of radius r. Hence flow in the vicinity of a high-pressure center must pass clockwise and with a net flow outward. If a more complete three-dimensional analysis of air flow near a high-pressure center is made, it is found that the flow is downward flow as well as outward. As a result, air must be brought in from above the center and settle downward to maintain the outward flow. This downward flow is called *subsidence* and is a possible inhibitor pollutant dispersion in the atmosphere.

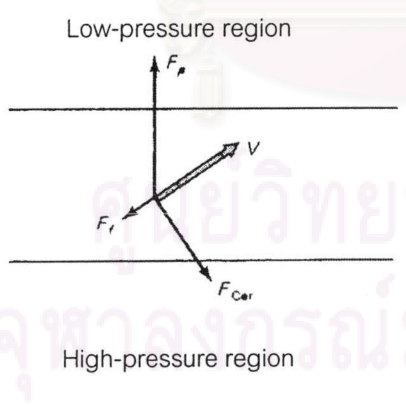

Figure 3.4 Effect of frictional force in the planetary boundary layer on the wind direction (Kenneth Walk, 1981)

Figure 3.5 (b) illustrates flow around a low-pressure center. In this case, the wind velocity vector is pointed in ward. Thus flow close to a low-pressure center is counterclockwise. Additional analysis reveals that the spiraling flow is upward as well as inward. Hence pollutants in the lower atmosphere will be carried upward and generally will be dispersed over a wide area. In addition, as the air rises it will cool as a result of the decrease in pressure at higher elevations. Water may then condense within the rising air mass. This effect may also have a cleansing action on a polluted atmosphere.

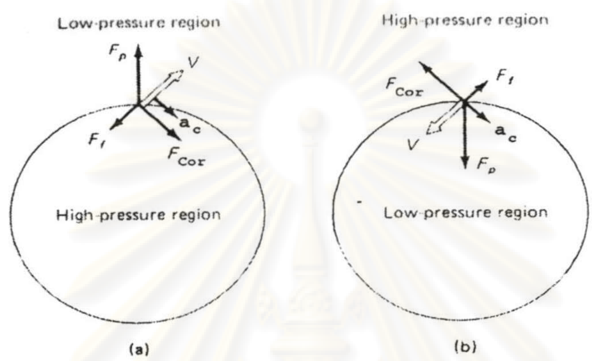

Figure 3.5 Force balances in the vicinity of high- and low-pressure regions. (a) Flow is outward and clockwise around a high-pressure region. (b) Flow is inward and counterclockwise around a low-pressure region. (Kenneth Walk, 1981)

3.3. Local effects

3.3.1 The Atmospheric Stability

One of the most important characteristic of the atmosphere is its stability. That is, its tendency to resist vertical motion or to suppress existing turbulence. This tendency directly influences the ability of the atmosphere to disperse pollutants emitted into it from natural or man-made sources. When a small volume of air is displaced upward in the atmosphere, it will encounter a lower pressure and undergo an expansion to a lower temperature. Usually the expansion is rapid enough that we can assume no heat transfer takes place between that parcel of air and the surrounding atmosphere. The change in temperature with elevation that is due to the adiabatic expansion is determined in the following manner.

The atmosphere is considered to be a stationary column of air in a gravitational field, and the air is approximated as a dry ideal gas. In the absence of frictional and inertial effects, a static force balance on a differential element of thickness dz leads to

$$dP = -\rho g dz \tag{3.1}$$

Where P is the atmospheric pressure, ρ is the atmospheric density (assumed to be constant), g is the local gravitational acceleration, and z is the elevation.

For an adiabatic process $dq = 0$; thus the first law of thermodynamics for a closed system containing an ideal gas undergoing a quasi-static change of state can be written as

$$C_p dT = \frac{1}{\rho} dP \tag{3.2}$$

Substitution of Equation (3.2) into Equation (3.1) and rearrangement gives

$$\left(-\frac{dT}{dz}\right)_{adia} = \frac{g}{c_p} \tag{3.3}$$

In SI units C_p for dry air equals 1.005 kJ/kg*°C at room temperature and g is 9.806 m/s^2 From these data one ascertains that for dry air

$$\left(\frac{dT}{dz}\right)_{dry\ adia} = -0.0098\ °C/m \quad = -0.0054\ °F/ft$$

It is convenient to define a *lapse rate* as the negative of the temperature gradient in the atmosphere. Thus the dry adiabatic lapse rate, which is given the special symbol Γ, is

$$\Gamma = \left(\frac{-dT}{dz}\right)_{dry\ \ adia} = 9.7\ °C/km \qquad = 5.4\ °F/1{,}000ft \tag{3.4}$$

The dry adiabatic lapse rate is extremely important in meteorological studies. The atmospheric stability can be principally determined from the lapse rate sketched in Figure 3.6

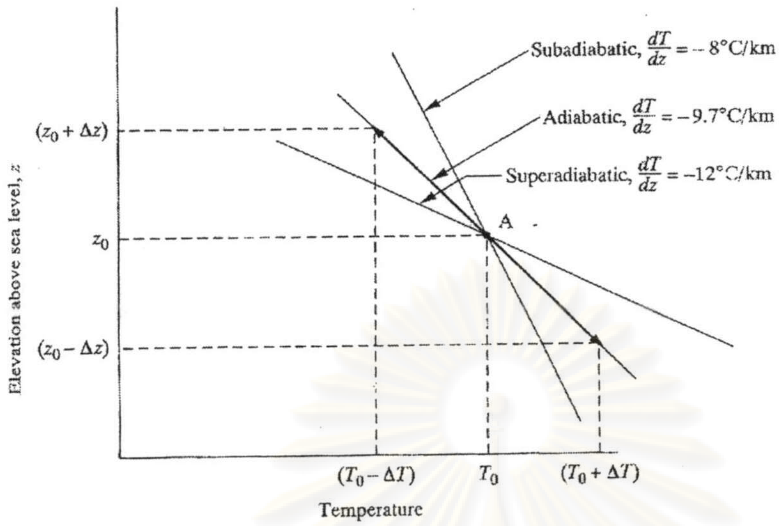

Figure 3.6 Behavior of air parcel displaced up or down in an atmosphere with an adiabatic, a subadiabatic, and a superadiabatic lapse rate. (Noel De Nevers, 1995)

If the surrounding air has the adiabatic lapse rate, then, no matter if the parcel is moved up or down, it will reach the same temperature as the surrounding air at its new location. In any such displacement the pressures of the displaced air and surrounding air will be practically equal; if they were not, the higher-pressure parcel would expand rapidly against the lower-pressure one. As a result, it will be at the same temperature and density as the surrounding air, and after this displacement gravity will not try to move it up or down. This condition is described as *neutral stability*: the air mass is neither stable nor unstable. In neutral conditions, which generally occur for moderate to high wind speeds, a vertically displaced parcel of air will neither rise nor fall any further. Such conditions thus result in strong mechanical mixing with negligible convective effects.

If the surrounding air has a *subadiabatic* lapse rate, the displaced parcel of air will have a different fate. If it is displaced upward, it will continue to follow the adiabatic curve, because its motion will be adiabatic, so that it will be colder and more dense than the surrounding air. Gravity wili drive it back down. If it is displaced downward, it will be warmer and less dense than the surrounding air, so gravity will drive it back up. Whatever disturbs its location, gravity will move it back toward its original location. Consequently, the condition is stable stability and vertical air motion is inhibited. Furthermore, turbulence is thus suppressed, and reduced mixing occurs.

If the surrounding air has a *superadiabatic* lapse rate, then if the parcel of air is displaced upward it will follow the adiabatic curve, so it will be warmer and less dense than the surrounding air. Thus gravity will drive it further upward. If it is displaced downward it will be colder and more dense than the surrounding air, so that gravity will drive it further downward. Whatever disturbs it from its original location, gravity will continue to move it in the direction of that displacement, and with an increasing velocity. Thus if the lapse rate is superadiabatic, the condition is unstable and vertical air motion is spontaneous if any minute disturbance triggers it. In very unstable condition, any vertically displaced air continues its movement, thus large convective cells are set and both turbulence and the consequent mixing are enhanced.

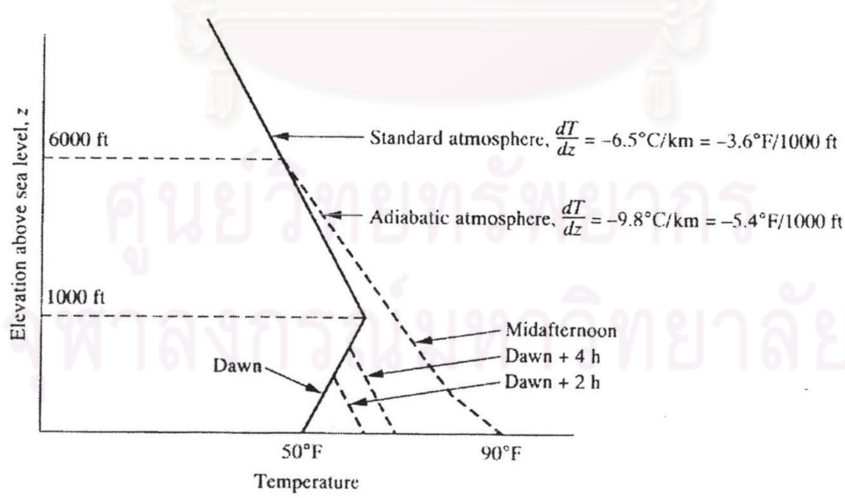

Figure 3.7 Vertical temperature distribution at various times on a cloudless day with low or average winds in a dry climate. (Noel De Nevers, 1995)

The all three stability conditions can occur at the same place but different times of day and on any clear dry, sunny day with low or average winds as shown in Figure 3.7. All night the ground surface has been cooling, and at dawn its temperature is perhaps 50°F. At infrared wavelengths the ground is an almost perfect blackbody radiator, so it is quite efficient at radiating heat to outer space. The ground surface has also been cooling the layer of air above it. The cooled air layer nearest the ground cools the layer of air above it, so that there is a steady flow of heat downward from the air to the ground by conduction, slight convection, and radiation. (Dry air, which is practically transparent to visible light, is not transparent to infrared radiation and does transfer some heat by infrared radiation.) At dawn, air temperature increases with elevation up to perhaps 1000ft. At that point the "cooling wave" from the ground runs into the lapse rate left over from the previous day, and the air temperature continues along up the standard atmosphere curve.

Below 1000 feet the air temperature at dawn increases with height. This pattern is called an *inversion*; such inversions occur every clear or slightly cloudy night, with low or average winds, on most of the world's land surface. Inside the inversion the situation is extremely stable; vertical disturbances are strongly damped out. Above the inversion, in the region with the standard lapse rate, the situation is mildly stable, vertical disturbances are damped, but not nearly as strongly as inside the inversion. This kind of inversion is the most common one and is called a *radiation inversion.*

When the sun comes up, it heats the ground surface, which heats the layer of air above it, by conduction, convection, and radiation. That layer heats the next layer above it, and so on. Two hours after dawn the ground temperature will be perhaps 70 °F. There will be a layer of warmed air near the ground, in which the lapse rate is practically the adiabatic lapse rate. At its top, this layer encounters the remainder of the previous night's inversion. Rising air from below cannot penetrate that inversion, for stability reasons. But at the boundary it mixes with the inversion, slowly destroying it, so that by four hours after sunrise the warmed air layer has grown and almost eliminated the inversion. To simplify the evaluation of the stability, there are many stability classification schemes. One of the famous stability classification schemes is Pasquill-Gifford's whose detail are the following:

3.3.1.1 Pasquill-Gifford Stability Classification

In this classification it is assumed that stability in the layers near the ground is dependent on net radiation as an indication of convective eddies and on wind speed as an indication of mechanical eddies. Insolation (incoming radiation) without clouds during the day is dependent on solar altitude, which is a function of latitude, day of the year, and time of day. When clouds are present, the extent of their coverage and thickness decreases the incoming and outgoing radiation. Daytime insolation must be modified according to the existing cloud cover. Six stability categories are defined in Table 3.1

Table 3.1 Pasquill-Gifford Stability Categories

Surface Wind (Measured at 10m)		Day-Time Insolation			Night-Time Cloudiness	
(m/sec)	(mph)	Strong	Moderate	Slight	Thinly Overcast or ≥ 4/8 Cloudiness*	≤ 3/8 Cloudiness*
<2	4.5	A	A-B	B	-	-
2-3	4.5-6.7	A-B	B	C	E	F
3-5	6.7-11.2	B	B-C	C	D	E
5-6	11.2-13.4	C	C-D	D	D	D
>6	13.4	C	D	D	D	D

*The degree of cloudiness is defined as that fraction of sky above the local apparent horizon that is covered by clouds.

Notes:

1. *Insolation* is the rate of radiation from the sun received per unit of earth's surface.
2. *Strong* insolation corresponds to sunny mid-day in summer. Slight insolation corresponds to similar conditions in mid-winter.
3. For A-B, B-C, etc. take the average of A and B values, respectively
4. Night refers to the period from 1 hour before sunset to 1 hour after dawn.
5. Regardless of wind speed, the neutral category D should be assumes for overcast condition during day or night and for any sky conditions during the hour preceding or following night.

1 mph (mile per hour) =0.4470 m/sec

A = Extremely Unstable	C = Slightly Unstable	E = Slightly Stable
B = Moderately Unstable	D = Neutral	F = Moderately Stable

3.3.2 **Mixing Height**

Figure 3.7 is also an illustration of a key concept in air pollution meteorology, the mixing height . In that figure, for the mid afternoon condition, there will be vigorous vertical mixing from the ground to about 6000 ft, and then negligible mixing above that height. The rising air columns that provide good vertical mixing induce large-scale turbulence in the atmosphere. This turbulence is three-dimensional, so it also provides good horizontal mixing. Pollutants released at ground level will be mixed almost uniformly up to the mixing height, but not above it. Thus the mixing height sets the upper limit to dispersion of atmospheric pollutants.

In the same figure we can see that in the morning the mixing height must be much lower and that it grows during the day. Similarly, we would except that the mixing height would be larger in the summer than the winter (see table 3.2.2)

Table 3.2 Typical values of the mixing height for the contiguous United States

	Mixing height, m	
	Range	Average
Summer morning	200-1100	450
Summer afternoon	600-4000	2100
Winter morning Winter	200-900	470
Winter afternoon	600-1400	970

In a major industrial city on a clear summer day, above the mixing height the air is clear and blue, whereas below the mixing height the air is hazy and brown or gray. The same phenomenon can be observed from mountains, hills and tall buildings when the mixing layer is thin.

3.3.2.1 Measurements of mixing depth

Before discussing measurements of mixing depth, it is necessary to define, as precisely as possible, what we are talking about. Nieustadt and van Dop define the top of the layer as follows:

(a) For unstable condition, the height of the convective boundary layer (CBL) is the height of the lowest temperature inversion

(b) For stable conditions, the height of the stable boundary layer (SBL) is the height at which the turbulence falls to 'say 5 percent' of the turbulence level 'near the surface'

In fact, mixing is occurring in the stable air at the top of the CBL and fluctuations in velocity due to gravity waves occur above the height at which turbulent mixing is insignificant in the stable boundary layer. Therefore neither of these definitions is entirely satisfactory. It is the maximum height from which material currently being released into the atmosphere with no buoyancy or vertical momentum will diffuse to the surface. It is also the maximum height which material released in a similar fashion from the surface will reach.

From the above it appears that there is no very precise definition and consequently no prospect of very precise measurement of the mixing depth.

The mid-point of the region where there is an obvious transition from the properties of the air mass in the mixing layer to the lower tropospheric air mass, is probably the definition. In many cases this will also correspond to a minimum in the wind velocity gradient and a maximum in the potential temperature gradient.

3.3.3 Winds

3.3.3.1 Wind velocity profile

Wind speed increases with elevation, in most of the troposphere. As we mentioned previously, the movement of air near the earth's surface is retarded by the frictional effects proportional to the surface roughness. Thus the nature of the terrain, the location and density of trees, the location and size of lakes, rivers, hills, and buildings produce different wind velocity gradients in the vertical direction. The air layer that is influenced by friction extends from a few hundred meters to several kilometers above the surface of the earth. Typically the wind will reach its frictionless velocity (called geostrophic or gradient velocity) at about 500 m (1640 ft) above ground. The region below this elevation, where ground friction plays a significant role, is the *planetary boundary layer*. The ground-level wind velocity is largely determined by how well this layer is coupled to the fast-moving geostrophic layer above it. When the atmosphere is stable or has an inversion, there is little vertical movement; and the coupling between the planetary boundary layer and the geostrophic wind is weak. Thus, inversions and stable atmospheres are normally associated with low ground-level wind velocity. The depth of this boundary layer is greater for unstable conditions than for stable conditions. Thus pollutants will be dispersed over a greater vertical distance under unstable atmospheric conditions.

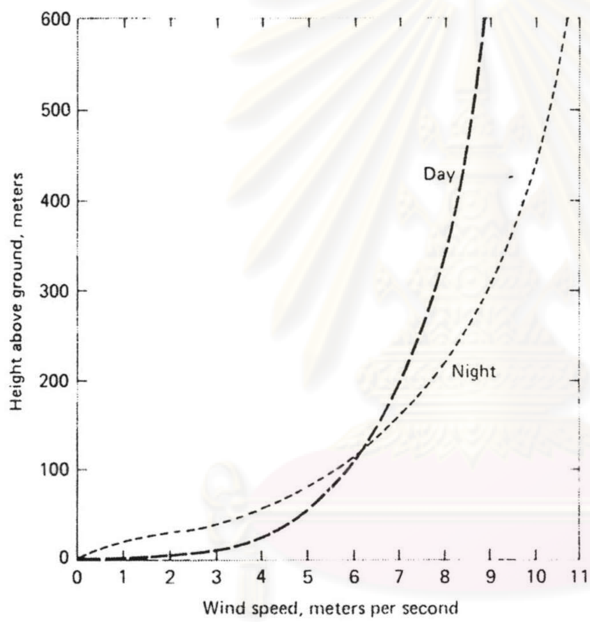

Figure 3.8 Change of wind-speed profile with stability.(source: D.B. Turner. *Workbook for Atmospheric Dispersion Estimates*. Washington, D.C.: HEW, 1969.)

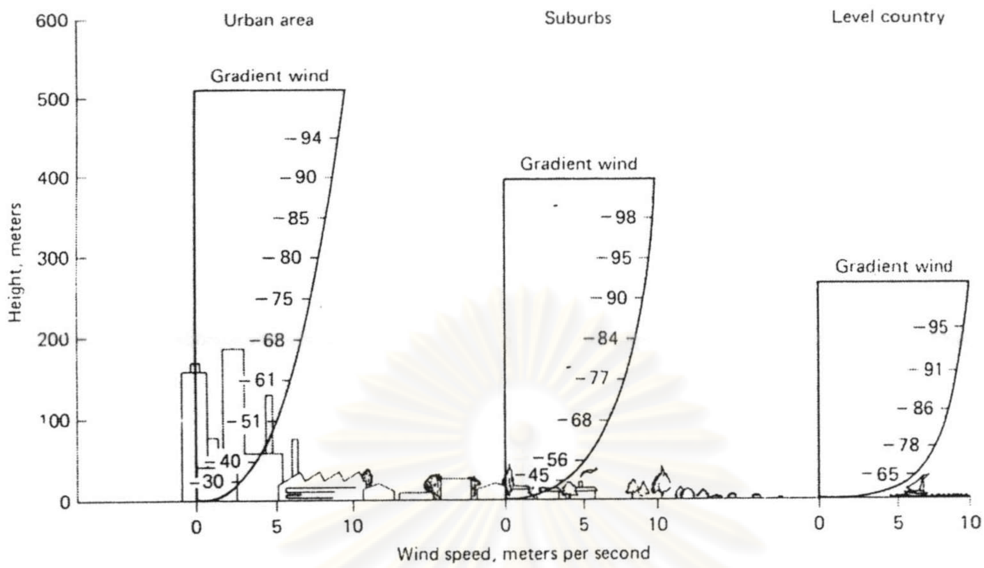

Figure 3.9 Effect of terrain roughness on the wind speed profile. With decreasing roughness, the depth of the affected layer becomes shallower and the profile steeper.(Davenport, 1963.)

Typical wind-speed profiles during the day and the night are shown in Figure 3.8. Because of a more stable atmospheric condition at night, the profile for the night is usually steeper than that for the day. Note that the wind-speed variation levels out in this case at an approximate height of 600 m. Above this height the frictional effect is negligible, and the wind speed becomes that of the gradient wind discussed earlier. The gross effect of terrain roughness on the wind-speed profile is shown in Figure 3.9 In this particular example the change in the overall boundary layer thickness is from approximately 500 m to 280 m, for decreasing roughness the profile also is steeper near the surface. Because of the appreciable change in wind-speed value must be quoted with respect to the elevation at which it was measured. The international standard height for surface wind measurements is 10 m. The wind direction and wind speed data at 10 m of Saraburi area during Jan-March, 2000are shown in Figure 3.10.

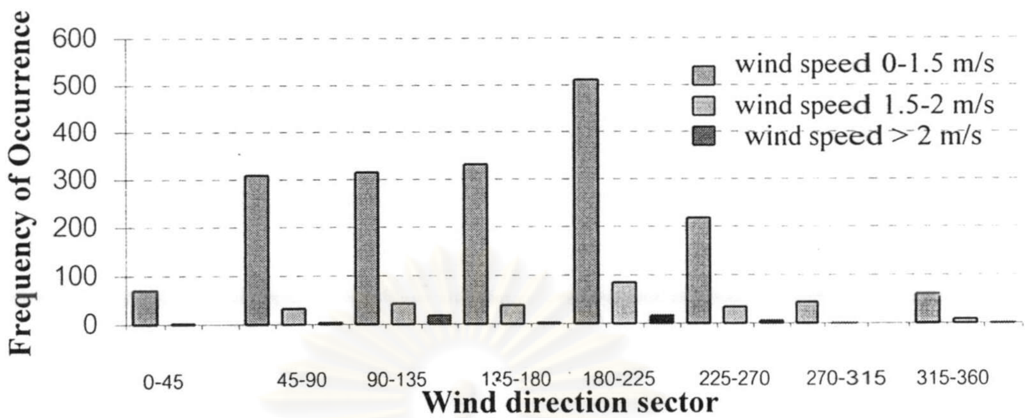

Figure 3.10 Wind speed and direction of Saraburi area during Jan-March, 2000 (source: Pollution Control Department, Bangkok, Thailand)

Frequently knowledge of the wind speed at some height other than at the standard is necessary. Numerous efforts have been made to develop suitable analytical expression relating wind speed to height. Because of the complexity of the phenomena, no completely satisfactory expression is available at this time. However, the power law for wind profile has been found useful for boundary layers up to several hundred meters in depth. Its details are as follows:

The Power Law for the Wind Profile: It is a common engineering practice to describe the wind profile with a power law

$$\frac{\bar{u}}{\bar{u}_m} = \left(\frac{z}{z_m}\right)^p \qquad (3.5)$$

Where $\bar{U}$ is the average wind speed at height z, $\bar{u}_m$ average wind speed measured at height z_m, and p is an exponent whose value is dependent upon atmospheric stability condition and surface roughness. Values of p have been found that range from 0.02 to 0.87. (Schnelle, K. B, 2000)

When the environmental lapse rate is approximately the adiabatic value and the terrain is generally level with little surface cover, the value of p to be chosen is approximately $\frac{1}{7}$.

The boundary layer thickness, and thus the wind-speed profile, is a function of the atmospheric stability. Hence the exponent p must vary in relation to the stability characteristics of the atmosphere, as a first approximation. Sutton (1953) has suggested that p can be related to a parameter n which is a function of the stability of the atmosphere. This relationship is

$$p = \frac{n}{2-n} \tag{3.6}$$

The value of n for various stability conditions are given in Table 3.3. The relationship between the temperature difference and parameter p from elevations of 5 to 400 ft is presented in Table 3.4.

Table 3.3 Relationship between the stability parameter n in equation (3.6) and the stability condition of the atmosphere (Schnelle, K. B, 2000)

Stability Condition	n
Large lapse rate	0.20
Zero or small lapse rate	0.25
Moderate inversion	0.33
Large inversion	0.50

Table 3.4 Relationship between the temperature difference and the parameter p for air layers from 5 to 400 ft thick, where $\Delta T = T_{400} - T_5$

$\Delta T(°F)$	p	$\Delta T(°F)$	p
-4 to –2	0.145	2 to 4	0.44
-2.5 to –1.5	0.17	4 to 6	0.53
-2 to 0	0.25	6 to 8	0.63
-1 to 1	0.29	8 to 10	0.72
0 to 2	0.32	10 to 12	0.77

According to the Pasquill-Gifford classes, the effect of surface conditions for urban and rural areas and stability on p values are given in Table 3.5.

Table 3.5 Estimates of the exponent Power p in velocity profile equation

	Stability Class					
	A	B	C	D	E	F
Urban	0.15	0.15	0.20	0.25	0.30	0.30
Rural	0.07	0.07	0.10	0.15	0.35	0.55

3.3.3.2 Wind direction

Superimposed on the wind circulation shown in Fig.3.11 are a series of disturbances call high-pressure zones (anticyclones) and low-pressure zone (cyclones). They are formed from large-scale instabilities, often involving the boundaries between the three circulation zones in each hemisphere.

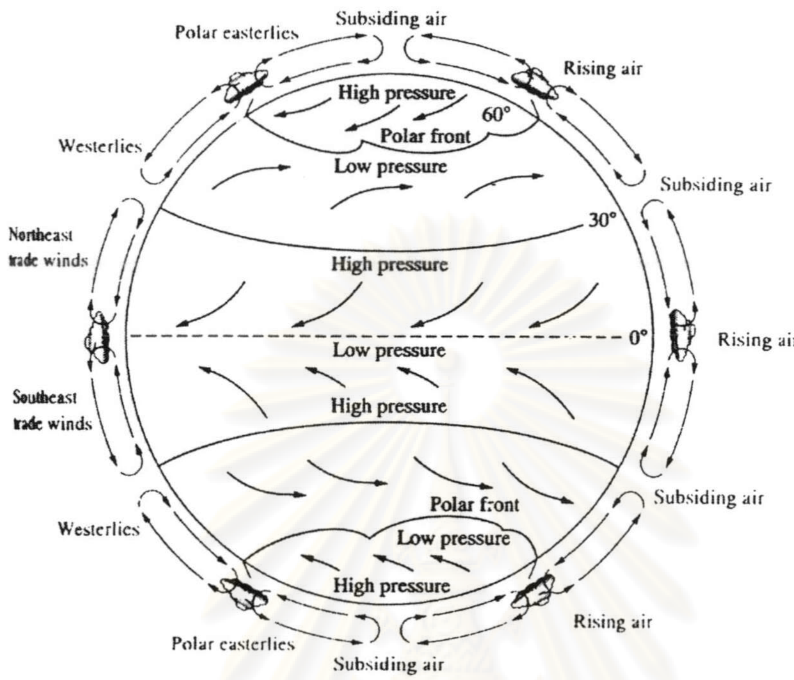

Figure 3.11 The wind circulation of the atmosphere (Frederick K. Lutgens/ Edward J. Tarbuck, the Atmosphere, 1992, p. 170.)

Near the center of a low-pressure area the winds associated with them can be strong enough to overwhelm all of the local effects. Mountains, valley, and shoreline all influence wind direction and magnitude as well as other meteorological parameters. The further details of topographical effects are discussed in the next section. Thus, estimating the wind direction at any time and any location can use the following rules of thumb:

1. Major, rapidly moving storms and fronts overwhelm local influence; local ground-level winds blow the way that the major storms dictate.

2. In deep valleys the daily alternation---wind up the valley in the daytime, down at night--- overcomes most other influences and determines most of the local flow when no major storm or frontal passage dominates. The valley effect is

greater in deep valleys than in shallow, in steep valleys than in gentle ones, at night than the daytime, and under conditions of light wind and clear sky than of strong wind or cloudiness.

3. On shore and offshore breezes dominate when there is no major storm. They are more likely to control the wind direction in light wind, clear sky conditions.

4. Absent all of the preceding or any other effects of local topography, the wind direction is more likely as shown in Fig.3.12 than any other. Figure 3.12 is a better predictor near the equator than near the poles.

3.3.4 The effect of the topography

Obstacles which project into the air stream produce additional turbulence and frictional effects. The low wind speeds which are being considered within this review will generally occur when the large-scale wind-forcing mechanism, in form of pressure gradients, are rather weak. In such case, local effects become significant; sea breezes, mountains and valleys etc. If the air is very stable, with little convection, the sea breeze will remain localized at the coast.

Local air flow patterns are greatly influenced by juxtaposed geographic configurations. For example: mountains and valleys, a mountain pass, or a land-water-air interface. The resulting meteorological changes are often described in terms of one variable, but a change in one variable must always be associated with changes in the others. However, the resultant weather features remain localized and do not travel in space.

3.3.4.1 The sea breeze: One common phenomenon of this type is the sea breeze which develops during the warm season along oceanic coastlines. A similar type breeze will occur at any land-air-water interface where a significant temperature difference exists between the land and water. During the day the water warms more slowly than the land. Air over the land warms and rises drawing in the cool air from over the water to replace the rising warm air. The rising warm air moves out to sea aloft where it cools and sinks. In this way a circulation cell develops from the water to the land at low levels and from land to sea aloft. The cell system is usually rather shallow, less than 1000 m high, but it may penetrate deep over the land, up to 40 to 80 km. Frequently at night after the sea breeze has dissipated, a land breeze will form where the circulation is weaker than the sea breeze and is more pronounced in the winter.

3.3.4.2 Flow pattern due to topographical feature: There is an infinite variety of flow patterns that can result from topographical or manmade obstacles. Vortices rotating in the vertical plane are formed when air flows over an obstacle. Separation of the flow may occur. Smoke plumes can be caught in these vortices and forced down to the ground. On the large scale, mountain ranges which produce uplifting of the air may produce large variations in rainfall throughout an extended geographic area. The Olympic and Cascade mountains are responsible for unusual rainfall distribution over the state. Figure 3.12 illustrates the distortions of the wind that can occur when the wind passes over a valley or through a mountain pass.

An isolated hill may affect the wind speed by causing a speedup of flow at the brow, with corresponding speed reductions upwind and downwind. In strong stable stratification, air is likely to flow around rather than over an isolated 3D hill, and would tend to be channeled along the axis of 2D obstructions. Whilst these flow feature may cause some effect on wind speed, with possible slight increases, the greatest effect would be on wind direction. Furthermore, mountains can act as barriers to low-level winds. For example, the Los Angeles Basin has high mountains on its north, east, and southeast. These impede the wind. Air mass are trapped in the basin and dilution of emission is prevented. As for Saraburi's topography, the area is quite mountainous as shown in Figure 3.13.

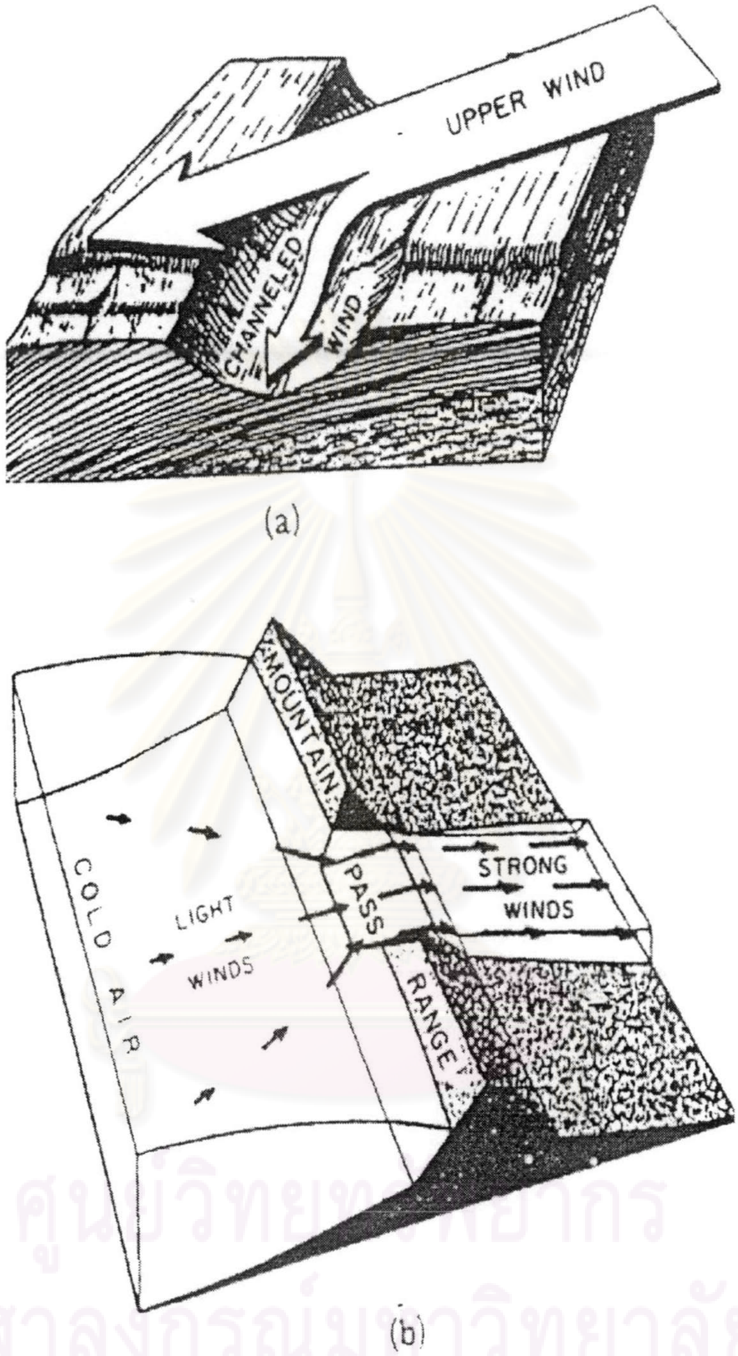

Figure 3.12 Distortions of the wind flow by topographic obstacles.(a) Channeling of the wind by a valley. (b) The effect of a mountain pass on the wind flow. (K. B. Schnelle, 2000)

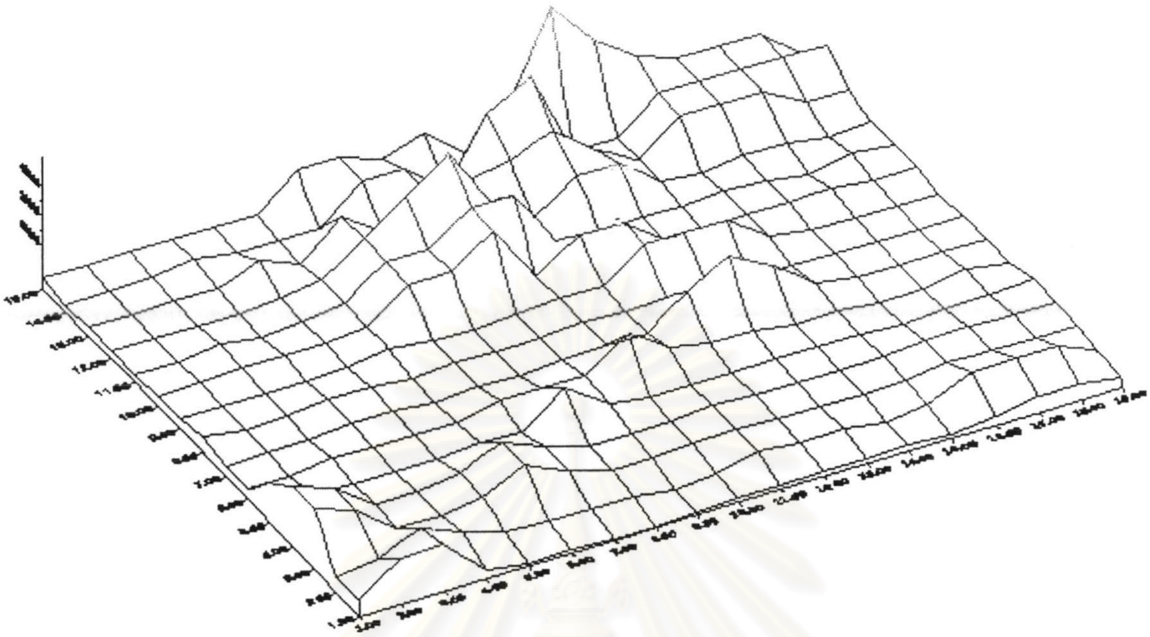

Figure 3.13 Topography of area of interest in Saraburi province. (Meechumna, P. 1999)

3.4 General characteristics of stack plumes

As noted in the preceding section, the dispersion of pollutants in the atmosphere is accomplished by two general mechanisms: the average wind speed and the atmospheric turbulence. The effect of the former is simply to carry the pollutants downwind the source; the latter causes the pollutants to fluctuate from the mainstream concentration in the vertical and crosswind directions. The two types of atmosphere turbulence—mechanical and convective—usually occur simultaneously under any atmospheric condition, but in varying ratios to each other. Because of these variations, the general geometric forms of gas plumes emitted from stacks are quite different.

Six classifications of plume behavior are noted in Figure 3.14. In addition to the general variation in geometric shape in the x-y coordinate plane, approximate velocity and temperature profiles are also shown. The *looping* plume shown in Figure 3.14(a) occurs when a high degree of convective turbulence exists. As noted on the figure, the looping plume indicates a superadiabatic lapse rate in the atmosphere, which leads to strong instabilities. A looping plume is usually associated with a clear daytime condition accompanied by strong solar heating of the earth's surface and light winds.

A coning plume [Figure 3.14(b)] occurs under essentially neutral atmospheric stability, and small-scale mechanical turbulence dominates. The coning plume occurs when skies are overcast during either the day or night. Winds are typically moderate to strong. The cloud cover prevents incoming solar radiation during the day and outgoing terrestrial radiation at night. Unlike in looping, in coning the major part of the pollutant concentration is carried fairly far downwind before reaching ground level in significant amounts. This is an especially good condition for estimating pollutant dispersion by the diffusion equations

A *fanning* plume occurs in the presence of large negative lapse rates, so that a strong surface inversion takes place to a considerable horizontal distance above the stack height. The atmosphere is extremely stable, and mechanical turbulence is suppressed. If the density of the plume is not significantly different from that of the surrounding atmosphere, the plume travels downwind at approximately constant elevation, as shown in Figure 3.14(c). As noted earlier, inversions are characteristic of clear nighttime conditions when the earth is cooled by outgoing radiation. When viewed from above, a *fanning* plume may appear to meander in the horizontal direction as it is carried downwind. It is difficult to predict downwind pollutant concentration and little pollutant effluent reaches the ground.

Fumigation plumes occur when a stable layer of air lies a short distance above the release point of the plume and an unstable air layer lies below the plume. Figure 3.14(d) illustrates the general profiles when an inversion exists aloft. The temperature profile necessary for fumigation usually arises in early morning, following a night characterized by a stable inversion. During that time period, relatively high ground-level concentrations will be reached. Fortunately the conditions for fumigation normally do not last for more than half an hour.

The conditions for the lofting plume indicated in Figure 3.14(e) are the inverse of those for the fumigation plume. The inversion layer lies below and the unstable layer lies through and above the plume. This is a favorable situation, since the pollutants are dispersed downwind without any significant ground-level concentrations. While fumigation conditions characterize the early morning after sunrise, lofting conditions prevail in the late afternoon and early evening under clear skies. During the day a negative temperature gradient is set up throughout the lower atmosphere, as a result of solar heating. In the late afternoon radiation from the surface leads to an inversion layer near ground level. As the inversion layer deepens, a lofting plume will change to a fanning plume. Hence lofting is usually a transitional situation. When an inversion exists both below and above stack height, *trapping* results. The diffusion of pollutants is severely restricted to the layer between the two stable regions, as shown in Figure 3-14(f).

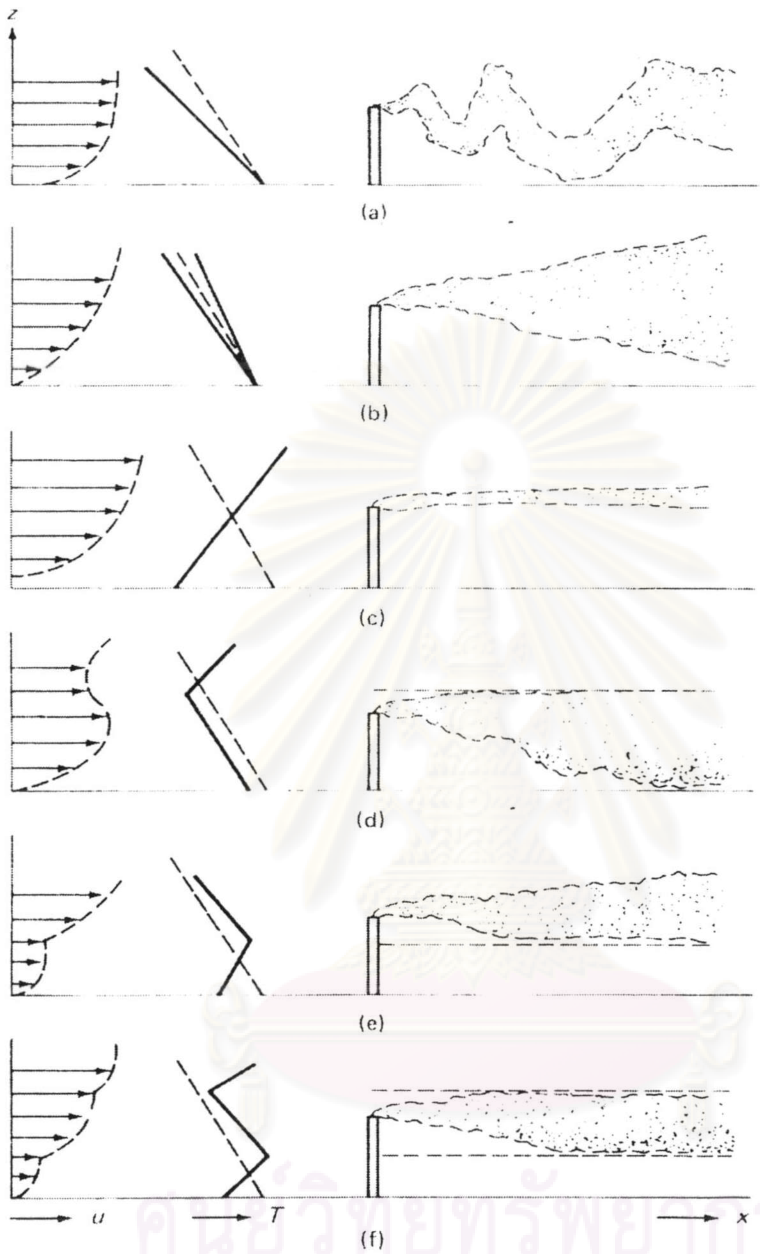

Figure 3.14 Typical velocity profile, temperature profile, and plume shape in the x-y coordinate system for various atmospheric conditions. (Dry adiabatic lapse rate, ---; ambient lapse rate,—.) (a) looping, strong instability; (b) coning, near neutral stability; (c) fanning, surface inversion; (d) fumigation, aloft inversion; (e) lofting, inversion below stack; (f) trapping, inversion below and above stack height. (Kenneth Walk, 1981)