CHAPTER 2

ATMOSPHERIC BOUNDARY LAYER

2.1 Introduction

The contents in this chapter are condensed and follow the well-established textbooks in field of atmospheric boundary layer study of Stull (1997), and Arya (2001). The purpose of this chapter is to introducing the background knowledge to the subject of atmospheric boundary layer as related to this research. It is intended for readers to be familiar with terms used in context of atmospheric boundary layer study and its physical meanings.

According to the vertical profile of temperature, the atmospheric structure is divided into four layers namely troposphere, mesosphere, and stratosphere. Troposphere is the lowest layer contains two sub layers, that is, the lower Atmospheric Boundary Layer (ABL) and the upper free atmosphere (Figure 2.1).

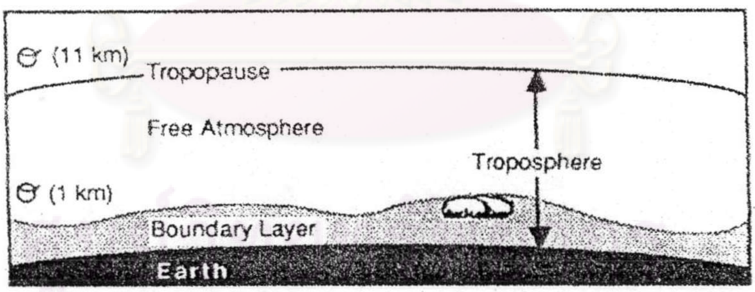

Figure 2.1 Troposphere composes of ABL (shaded) and free atmosphere. ABL closed to earth's surface and covered by free atmosphere. (Stull; 1997)

The daily cycle of radiative heating causes a daily cycle of sensible and latent heat fluxes between the earth and the air. However, these fluxes cannot directly reach the whole atmosphere. They are confined by the troposphere to a shallow layer near the surface that is the Atmospheric Boundary layer (ABL). Therefore ABL is directly influenced by the presence of the earth's surface, and responds to surface forcing with a timescale of about one hour or less Stull (1997), Arya (2001). The forces come from friction drag, evaporation and transpiration, heat transfer, pollutant emission and terrain induced flow modification. ABL is distinguished from free troposphere by temperature inversion which acts like a lid or cap to ABL motion by influence of the surface. The ABL thickness is variable in time and space from hundreds of meter to a few kilometers.

Understanding of ABL structure and ABL top height can examine the relationship between temperature and aerosol that interested in present study (Russel, 1974). ABL diurnal depth is an important variable in precipitation modification by urban area (Huff and Changnon, 1973) because they have show good correlation with the time and frequency of convective cloud development (Spangler and Dirks, 1974) and the precipitation anomaly (Beebe and Morgan, 1972). In addition, during the early morning and other period of stability the mixing depth is important to are pollution, since it is one of the two principal variation (the other is wind) that govern pollutant dilution over moderately long distances.

In addition to dynamics of the mixing depth (include the behavior of convective plumes), other boundary layer processes are fundamental interest for the formation of convective clouds and associated air motions, the behavior and effects of large scale subsidence, the transport and mixing of neighboring air mass, and the generation and propagation of atmospheric wave.

2.2 ABL Evolution and Structure

For most obvious cases, the ABL can be distinguished from free troposphere by temperature inversion which acts like a lid or cap to ABL motion by influence of the surface. ABL's characters, however, vary both in temporal and spatial.

For temporal variation, surface temperature induces the air at bottom of ABL to rise up and take place by downward air from top of ABL to form circulation. The convection causes mixing of air therefore meteorological variations such as potential temperature, atmospheric constituents become homogeneous. Other temporal variations of the ABL height and structure often occur as a result of the development and passage of meso and synoptic systems. Generally, the ABL becomes thinner under influence of a large scale subsidence (downward motion) and the low-level horizontal divergence associated with high pressure system (anticyclone). On the other hand, the ABL can grow to be very deep and merge with towering clouds in disturbed weather conditions that are associated with low-pressure system (cyclone) (Figure 2.2). In this case, it is difficult to distinguish the ABL top from in-cloud circulations; the cloud base is generally used as an arbitrary cutoff for the ABL top.

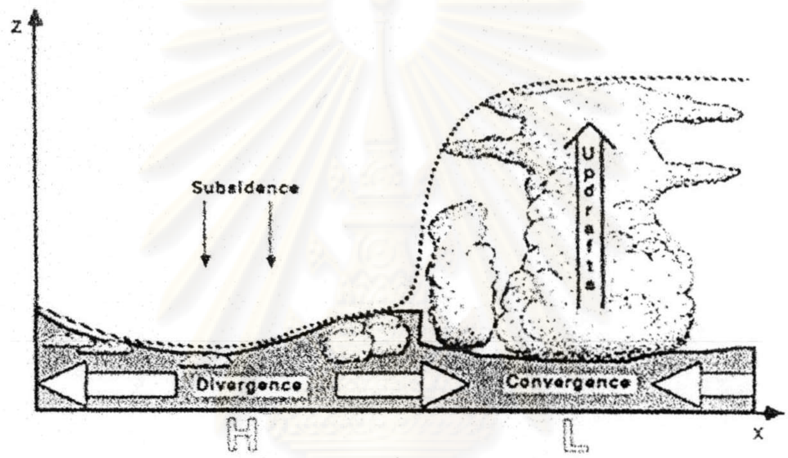

Figure 2.2 Influence of synoptic scale vertical circulation on the ABL. (Stull; 1997)

Spatial variations of the ABL depth and structure occur as a result of changes in land use and topography of the underlying surface.

ABL is often divided into 2 conditions as below;

2.2.1 Marine Boundary Layer

Small diurnal changes of the water surface temperature are due to the large heat capacity of the mixed layer in water. Therefore diurnal variation of ABL height and other meteorological variables are found to be much smaller in this area.

Most changes in ABL depths over oceans are caused by synoptic processes of vertical motion and advection of different air masses over the sea surface.

2.2.2 Terrestrial Boundary layer

Over the land, the ABL is influenced by strong solar heating at the surface. Diurnal variation is one of the dominant characteristics of the ABL over land. The diurnal variation is not formed by solar radiation directly but by the warming and cooling of ground by influence of solar radiation. delete Turbulence is sometime used to define the ABL (detail of turbulence will be discussed later in section 2.4).

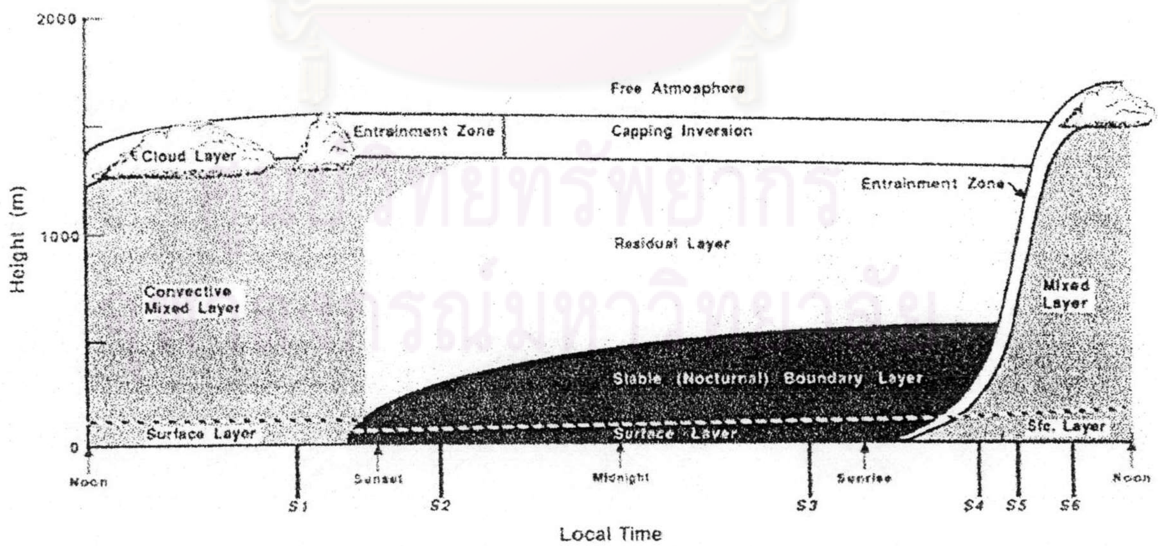

Figure 2.3 The ABL components at high pressure region over land surface. (Stull; 1997)

In low pressure regions the upward motions carry ABL air away from the earth's surface to large altitudes through out the troposphere. Therefore cloud base is often used as an arbitrary cut-off for ABL studies in these cases (Figure 2.2).

Over land surface in high-pressure region the ABL consists of three components, namely mixed layer, residual layer and stable boundary layer (Fig.2.3).

2.2.2.1 Mixed Layer (ML)

Turbulent nature of the ABL is one of its conspicuous and important features (Garratt, 2000). The solar radiation makes the surface warmer than the air and conduct heat energy to the air, then the air become unstable and thermal of warm air rises from the surface. This occurs typically in fair weather and cold air blows over warmer surface, during daytime. Under this condition ABL is called mixed layer. Turbulence causes mixing of the air at the top of ABL and the air at the bottom of ABL. The resulting of mixture is heat, moisture, and momentum in the mixing layer uniformly in vertical.

Air movement in mixed layer can express by pollutants emitted from smoke stacks which exhibit a characteristic looping as those portions of the effluent emitted into warm thermals begin to rise (Figure 2.4).

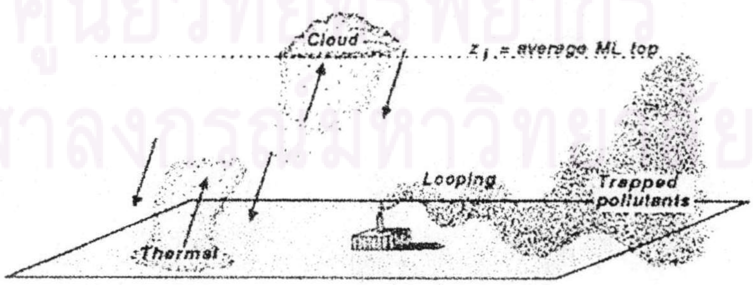

Figure 2.4 Idealization of thermals in a mixed layer. Smoke plumes loop up and down in the mixed layer eventually becoming uniformly distribution. (Stull; 1997)

Wild (1985) suggested growth of mixed layer depth composed of 4 phases (Figure 2.5). The description expressed as below;

1. During the early morning the mixing layer depth is shallow. The depth increases slowly because strong nocturnal stable layer makes caps to the mixing layer evolution.

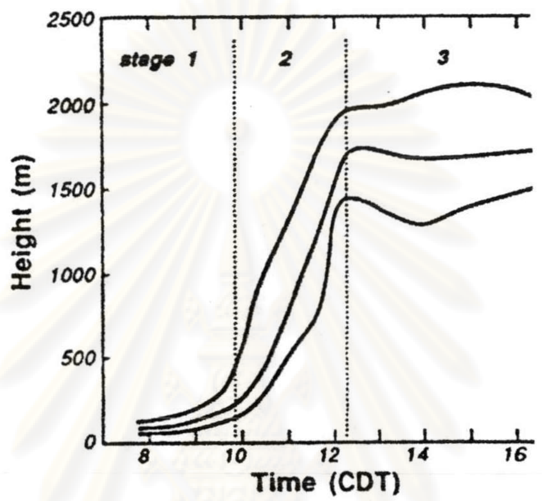

Figure 2.5 ABL modification by Wild (1985)

2. Late morning, the Mixing layer depth penetrates rapidly to the base of residual layer base due to effective solar radiation.
3. When the thermal reaches the capping inversion of the top of residual layer, the movement meets resistance and the Mixing layer growth rate rapidly decreases. During this phase the Mixing layer depth is almost constant.
4. As the sun set, the generation rate of convective turbulence decrease to the point where turbulence can no more be maintained. In the absence of mechanical forcing, turbulence in the Mixing layer decays completely, causing us to reclassify the layer as the residual layer.

2.2.2.2 Residual Layer (RL)

About half of hour before sunset the thermals forming is stopped and allows turbulence to decay in the formerly well mixed layer. The resulting layer of air is sometimes called the residual layer because its initially mean state variables and concentration variables are the same as those of the recently decayed mixed layer. Causing residual layer is neutrally stratified, resulting in turbulence that is nearly of equal intensity in all directions. Smoke plumes emitted into the residual layer tend to disperse at equal rates in the vertical and lateral direction, creating a cone shape plume. In addition, variables usually decrease slowly during the night because of radiation

2.2.2.3 Stable Boundary Layer (SBL)

After sunset the land surface become cooler than the air and the wind at surface is normally gentle in the nighttime. Therefore this layer is called stable boundary layer.
Pollutants emitted into the stable boundary layer disperse relatively in the vertical. They disperse more rapidly in horizontal. This behavior is called fanning.

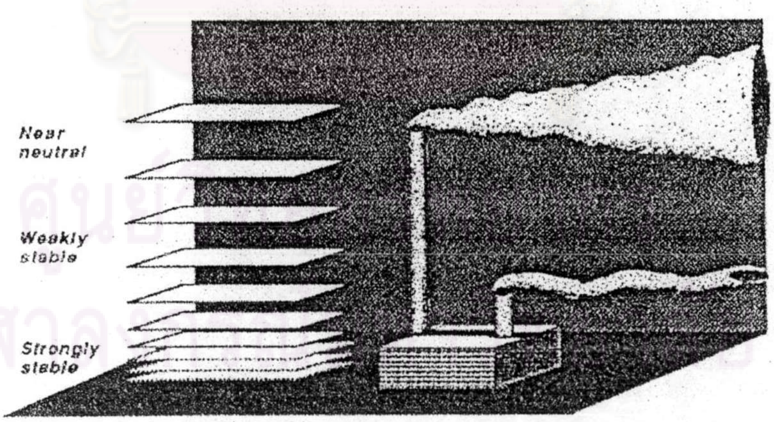

Figure 2.6 Pollutants emitted form cone shape plume in residual layer and faning in stable boundary layer. (Stull; 1997)

2.2.3 Turbulence and Mixing

The dynamics of the ABL is important for the transfer flux (F) of heat, momentum, water vapor and the other chemical species between free troposphere and earth surface. The rate of change of F through the air parcel is expressed as below;

$$\frac{dS}{dt} = \frac{-d}{dx}(F_x) - \frac{d}{dy}(F_y) - \frac{d}{dz}(F_z). \tag{1}$$

In the atmosphere, transfer of properties is conducted by

- advection
- conduction
- turbulence

2.2.3.1 Advection

Advections carry moisture, heat and the other atmospheric character from one place to the other place in the atmosphere. The atmospheric parameters transfer is like fluid transportation in a pipe. If the fluid velocities U and density ρ transport in a pipe had cross section A, then the flux F would be ρUA and average mass flux per unit area would be $\overline{\rho U}$. If the system is close system then quantity of fluid is conserved in the system. The flux of quantity S is defined as $\overline{SU}$. The rate of change of quantity S in the pipe can be expressed as below

$$\frac{d}{dt}(A \cdot dx \cdot S) = A(F_{in} - F_{out}) = A\left(\overline{SU}_{in} - \overline{SU}_{out}\right) \text{ or}$$

$$\frac{dS}{dt} = \frac{-d}{dx}(F) = \frac{-d}{dx}\left(\overline{SU}\right). \tag{2}$$

or in three dimensions as

$$\frac{d\rho}{dt} = \frac{-d}{dx}\left(\overline{\rho U}\right) - \frac{d}{dy}\left(\overline{\rho V}\right) - \frac{d}{dz}\left(\overline{\rho W}\right).$$

In advection, however, there will be no mean vertical velocity **W**, so a two dimensions representation is appropriate.

2.2.3.2 Conduction and Diffusion

The initial assumption is that any diffusive flux is proportion to the gradient of the quantity S

$$F_x \propto dS/dx. \tag{3}$$

The flux is actually against the gradient, as smoke moves from a high concentration to a low concentration. If we consider vertical gradients instead of horizontal, equation (3) becomes

$$\frac{dS}{dt} = \frac{-d}{dz}(F) = \frac{-d^2}{dz^2}(\alpha S). \tag{4}$$

For ABL, the surface fluxes are defined as diffusive fluxes.

At the earth's surface, the energy balance equation is simply expressed as

$$R_n = H + l_{vap}E + G.$$

R_n as the net radiative flux,

H as the sensible heat flux

l_{vap} E as latent heat flux and

G as the energy flux through the ground or

Considering the vertical diffusion through the ABL, sensible heat flux and latent heat flux may be defined as

$$H = -\rho c_p \kappa_n \left(\frac{dT}{dz} \right),$$

$$l_{vap} E = \rho l_{vap} d \left(\frac{dq}{dz} \right).$$

Here κ_n defines the diffusivity of heat ($\sim 2.4 \times 10^{-5}$ $m^2 s^{-1}$)

d defines the diffusivity of water vapor ($\sim 2.9 \times 10^{-5}$ $m^2 s^{-1}$)

The transfer of momentum is also of great value in atmospheric dynamics

$$\tau = shearing\ stress = \rho \nu \left(\frac{du}{dz} \right).$$

with ν define as the kinematics viscosity ($\sim 1.3 \times 10^{-5}$ $m^2 s^{-1}$).

For the atmosphere, the diffusivity/viscosity is relatively small. It is only effective over very short distances, typically in order of millimeters. When initially discussed the entire ABL, we defined the molecular layer at the lowest millimeters next to the earth surface.

2.2.3.3 Convection and Turbulence transport

Circulation or eddies effectively mixes fluid both up and down at the same time. The upward rotation would bring moisture or what ever from the surface upward while the downward part of the rotation from top of atmosphere would bring dry air or whatever downward. Turbulence is the composition of eddies. The interested is a range size of eddies which commonly referred to the spectrum of turbulence of kinetic energy. In three dimensions, there is a flow of energy from large eddies to the smallest eddies. Thus the large eddies essentially decay into smaller eddies. This process continues until diffusion removes the kinetic energy.

For turbulence, it would be easy to write the velocity and some other variable as a combination of the overall mean and the fluctuations or Raynolds decomposition.

$$w = \overline{W} + w' \ \ and\ \ s = \overline{S} + s' \tag{5}$$

We could consider transportation on the fine scale with the advection equation.

$$\frac{dS}{dt} = \frac{-d}{dz} ws.$$

We have to integrate over distance or time to make this meaningful. We define the mean turbulence flux as

$$F = \frac{1}{T}\int ws\ dt = \left\langle \overline{S \cdot W} \right\rangle + \left\langle \overline{S \cdot w'} \right\rangle + \left\langle \overline{s' \cdot W} \right\rangle + \left\langle \overline{s' \cdot w'} \right\rangle = \left\langle \overline{s' \cdot w'} \right\rangle \tag{6}$$

Here T is the length of the time interval. Our turbulence flux is simply $\langle s' \cdot w' \rangle$, the correlation between two fluctuations since $\overline{W}$ is zero.

2.3.3.1 Turbulence Diffusion

According to the first assumption, flux is proportion to the gradient, therefore turbulence diffusion may be expressed as below

$$\left\langle \overline{s' \cdot w'} \right\rangle \propto \frac{d}{dz} s$$

Then the turbulence diffusion has to be assumed much greater than the molecular level.

Another common approximation for turbulence flux is to be assumed that flux is proportional to the third order derivative. This is called hyper-diffusivity and can be lead to the equation

$$\frac{dS}{dt} = -\alpha \frac{d^2}{dz^2}\left(\frac{d^2}{dz^2} S\right) \tag{7}$$

2.3.3.2 Mixing Length Theory

The turbulence of the surface layer is predominantly driven mechanically by horizontal wind change as altitude change over the surface. The mixing through the surface layer can be accurately described by mixing length theory. At a given altitude, there is a dominant eddy size that called mixing length, l.

$$s' \propto l \frac{ds}{dz}, \tag{8}$$

The mixing length, l assumed it is proportional to height;

$$l = kz. \tag{9}$$

Thus mixing at high level is more effective than mixing near the surface. Field observations have strongly supported this theory. Further, these observations have led to an empirical value for *k* as 0.4, which is known as von Karman's constant.

Let's consider the momentum equation

$$\frac{\partial}{\partial t}(\rho u) = -\frac{\partial}{\partial z}\tau = -\frac{\partial}{\partial z}\left(\overline{\rho u' w'}\right).$$

If the turbulence is isotropic then $u' \sim w'$

$$\frac{\partial}{\partial t}(u) = -\frac{\partial}{\partial z}\left(\overline{l^2\left(\frac{\partial u}{\partial z}\right)^2}\right).$$

Further assumption the flow is steady, and then there is no time dependence.

$$\frac{\partial}{\partial z}\left(\overline{l^2\left(\frac{\partial u}{\partial z}\right)^2}\right)=0 \Rightarrow \left(\frac{\partial u}{\partial z}\right)=\frac{u^*}{kz},$$

$$u(z)=\frac{u^*}{k}\ln(z/z_0). \tag{10}$$

There are two free variables in this expression

u^* is the friction velocity and

z_0 is the surface roughness length which is measure of how rough of the surface such as 0.01 millimeter for ice or 100 meters for mountain.

2.2.4 Potential temperature

The ABL is often turbulence. Namely, within turbulence region, warmer potential temperature air from the top of ABL is mixed with cooler potential temperature air at the bottom of ABL. The result mixture has a medium potential temperature that is uniform with height.

The potential temperature θ can be defined that removes the effect of "dry" adiabatic temperature changes experienced by air parcel during vertical motion. If height is used as the vertical coordinate, then

$$\theta(z) = T(z) + \Gamma_d \cdot z$$

For pressure coordinate

$$\theta = T \cdot [P_0/P]^{\Re_{d}/C_p}$$

Where P_0 is a reference pressure

To include the buoyant effect of water vapor and liquid water in air, we can define a virtual potential temperature in non cloudy and cloudy air as below;

In non cloudy air $$\theta_v(z) = \theta \cdot (1 + 0.61 \cdot r)$$

$$\theta_v(z) = \theta \cdot (1 + 0.61 \cdot r_s - r_L)$$