The Atmospheric Boundary Layer

- Turbulence (9.1)
- The Surface Energy Balance (9.2)
- Vertical Structure (9.3)
- Evolution (9.4)
- Special Effects (9.5)
- *The Boundary Layer in Context* (9.6)
• Advances will continue to be made in boundary layer meteorology due to

• advances in measurement capabilities, both *in situ* and remote

• increased *computational* capabilities of research numerical models

• increases in resolution of *NWP* models

• needs of *wind power* generation, *urban* meteorology, *wildfire* science, etc
Relative backscatter intensity (dB) at 1543 nm

Other conditions being the same, alpine glaciers and snow-fields lose more mass on a humid summer day than on a dry summer day. On a clear, calm day, the surface sensible heat flux into the air does not usually become positive until 30 to 60 min after sunrise. In fair weather, the heat and momentum fluxes at the top of the mixed layer due to entrainment are usually downward, but the moisture flux is positive. You can estimate the static stability of the boundary layer by looking at the shape of the smoke plume from a smoke stack. In Fig. 9.40, why do the surface wind speed and the cloudiness increase as the air flows northward across the sharp front in the sea-surface temperature field that lies along 1°N?
Boundary Layer Winds

For the special case of horizontally homogeneous turbulence above the viscous sublayer (5.9), molecular viscosity and horizontal turbulent momentum flux divergence terms can be neglected. The mean flow momentum equations (5.9) and (5.10) become

\[
\frac{D\overline{u}}{Dt} = -\frac{1}{\rho_0} \frac{\partial \overline{p}}{\partial x} + f\overline{v} - \frac{\partial \overline{u}'\overline{w}'}{\partial z} \tag{5.16}
\]

and

\[
\frac{D\overline{v}}{Dt} = -\frac{1}{\rho_0} \frac{\partial \overline{p}}{\partial y} - f\overline{u} - \frac{\partial \overline{v}'\overline{w}'}{\partial z}. \tag{5.17}
\]

In the boundary layer the turbulent terms must be included, and the resulting approximation is

\[
f(\overline{v} - \overline{v}_g) - \frac{\partial \overline{u}'\overline{w}'}{\partial z} = 0, \tag{5.18}
\]

\[
-f(\overline{u} - \overline{u}_g) - \frac{\partial \overline{v}'\overline{w}'}{\partial z} = 0 \tag{5.19}
\]
a three-way balance between the Coriolis force, the pressure gradient force, and the turbulent momentum flux divergence. Here

\[
\mathbf{V}_g \equiv \mathbf{k} \times \frac{1}{\rho f} \nabla p \tag{2.23}
\]
is used to express the pressure gradient force in terms of geostrophic velocity.
Fig. 5.3 Balance of forces in the well-mixed planetary boundary layer: $P$ designates the pressure gradient force, $Co$ the Coriolis force, and $F_T$ the turbulent drag.
Boundary Layer Winds

5.3.1 Well–Mixed Boundary Layer

If a convective boundary layer is topped by a stable layer, turbulent mixing can lead to the formation of a well–mixed (ML) layer. Such boundary layers are common over land during the day when surface heating is strong and over oceans when the air near the sea surface is colder than the surface water temperature.

Observations also indicate that the surface momentum flux can be represented by a *bulk aerodynamic formula* where

\[
(u' w')_s = -C_d |\bar{V}| \bar{u}
\]

and

\[
(v' w')_s = -C_d |\bar{V}| \bar{v}.
\]

\(C_d\) is a nondimensional *drag coefficient*, \(|\bar{V}| = (\bar{u}^2 + \bar{v}^2)^{1/2}\), and the subscript \(s\) denotes surface values (taken at standard anemometer height). Observations show that \(C_d \sim 1.5 \times 10^{-5}\) over oceans, and is several times as large over land.
Boundary Layer Winds

Using this representation for surface momentum flux, (5.18) and (5.19) can be integrated from the surface to the top of the boundary layer where \( z = h \) to give

\[
f(\bar{v} - \bar{v}_g) = -\left(\frac{u'w'}{\bar{h}}\right)_s = \frac{C_d |\bar{V}| \bar{u}}{h},
\]

\[
-f(\bar{u} - \bar{u}_g) = -\left(\frac{v'w'}{\bar{h}}\right)_s = \frac{C_d |\bar{V}| \bar{v}}{h}.
\]

By choosing the axes such that \( \bar{v}_g = 0 \), (5.20) and (5.21) are written

\[
\bar{v} = \kappa_s |\bar{V}| \bar{u}, \quad \bar{u} = \bar{u}_g - \kappa_s |\bar{V}| \bar{v}
\]

where \( \kappa_s \equiv C_d/(fh) \). You have already seen this wind behavior in your Synoptic Meteorology class; near the surface winds tend to spiral into lower pressure centres, out of high pressure centres. In the mixed layer wind speed is less than geostrophic speed, and there is a component of motion directed toward lower pressure (that is, to the left of the geostrophic wind in the Northern Hemisphere and to the right of the geostrophic wind in the Southern Hemisphere) whose magnitude depends on \( \kappa_s \). If we plug in typical values for \( \bar{u}_g \) and \( \kappa_s \) into (5.22), \(|\bar{V}|\) will be less than the magnitude of the geostrophic wind.
Use the mixed layer expressions for $\bar{u}$ and $\bar{v}$ to calculate $\bar{u}$, $\bar{v}$, $|\vec{V}|$, and the cross-isobar angle of $\vec{V}$ for $\bar{u}_g = 10$ m/s and $\kappa_s = 0, 0.015, 0.05, 0.1, 0.2,$ and $0.4$ s/m.
or, in component form, satisfied:

\[
\frac{\partial \mathbf{V}}{\partial t} + \mathbf{V} \cdot \nabla \mathbf{V} = \frac{-1}{\rho} \nabla P + \mathbf{f} \times \mathbf{V} + \frac{1}{
\rho \Theta} \nabla \Theta \times \mathbf{V}
\]

Changes in velocity is on the order of a day or so which individual air parcels experience significant 10 m s\(^{-1}\) where wind field are on the order of 10 m s\(^{-1}\) and extratropical cyclones, typical horizontal velocities In large-scale wind systems such as baroclinic waves are very similar. Therefore, within about 10\% in middle latitudes, about an order of magnitude larger than the typical horizontal accelerations of air parcels.

Making use of the vector identity (7.14) it is defined on constant pressure surfaces so that

\[
\nabla \Phi = \frac{1}{\rho} \mathbf{f} \times \mathbf{V} + \frac{1}{
\rho \Theta} \nabla \Theta \times \mathbf{V}
\]

For any given horizontal distribution of pressure on the Earth and its relationship to the geostrophic wind, the pressure gradient force and hence, the potential height contours, pointing toward higher values. Thus a typical parcel acceleration in (7.15a) the horizontal wind field is defined on surfaces) it is possible to define a geostrophic wind field for which this relationship is exactly balanced the Coriolis force.

\[
\mathbf{V}_g = \frac{\nabla \Phi}{\nabla \Phi \cdot \mathbf{i}}
\]

\[
\frac{\partial \mathbf{V}_g}{\partial t} + \mathbf{V}_g \cdot \nabla \mathbf{V}_g = \frac{-1}{\rho} \nabla P + \mathbf{f} \times \mathbf{V}_g
\]

However, pressure surfaces are sufficiently flat that the geopotential surfaces (or geopotential height on pressure surfaces) are on the order of 10 m s\(^{-1}\). Thus, to within about 10\% in middle latitudes, the horizontal equation of motion is usually very small, the only term that is capable of overcoming the pressure gradient force.

From the Greek: geo = earth, strophic = circular, wind that circulates cyclonically around a center of low pressure and vice versa, as in Fig. 1.14, justifying the identification of local pressure minima with cyclones and high latitudes, the horizontal acceleration due to the pressure gradient force per unit mass 5 x 10\(^{-3}\) ms\(^{-2}\). Thus a typical parcel acceleration P is directed normal to the isobars (or geopotential height field). In the free atmosphere, where the frictional force is the scalar geostrophic wind speed and its relationship to the horizontal pressure gradient force or 10\%.

Fig. 7.9 illustrates this relationship and the timescale over which this relationship is exactly true. In Fig. 7.10, as in Fig. 7.9, low pressure maxima are represented in Fig. 7.9. As in Fig. 7.9, the stronger the Coriolis force required to balance the pressure gradient force and hence, the spacing of the isobars or geopotential height contours, pointing toward higher values. In either hemisphere, the geostrophic wind field is parallel to the isobars, leaving low pressure to the left. In either hemisphere, the geostrophic wind field circulates cyclonically around a center of low pressure.