Lecture 1

Scope of Boundary Layer (BL) Meteorology

In classical fluid dynamics, a boundary layer is the layer in a nearly inviscid fluid next to a surface in which frictional drag associated with that surface is significant (term introduced by Prandtl, 1905). Such boundary layers can be *laminar* or *turbulent*, and are often only mm thick.

In atmospheric science, a similar definition is useful. The *atmospheric boundary layer* (ABL, sometimes called P[lanetary] BL) is the layer of fluid directly above the Earth's surface in which significant fluxes of momentum, heat and/or moisture are carried by turbulent motions whose horizontal and vertical scales are on the order of the boundary layer depth, and whose circulation timescale is a few hours or less (Garratt, p. 1). A similar definition works for the ocean.

The complexity of this definition is due to several complications compared to classical aerodynamics.

i) Surface heat exchange can lead to thermal convection

ii) Moisture and effects on convection

iii) Earth's rotation

iv) Complex surface characteristics and topography.

BL is assumed to encompass surface-driven dry convection. Most workers (but not all) include shallow cumulus in BL, but deep precipitating cumuli are usually excluded from scope of BLM due to longer time for most air to recirculate back from clouds into contact with surface.

Air-surface exchange

BLM also traditionally includes the study of fluxes of heat, moisture and momentum between the atmosphere and the underlying surface, and how to characterize surfaces so as to predict these fluxes (roughness, thermal and moisture fluxes, radiative characteristics). Includes plant canopies as well as water, ice, snow, bare ground, etc.

Characteristics of ABL

The boundary layer itself exhibits dynamically distinct sublayers

i) Interfacial sublayer - in which molecular viscosity/diffusivity dominate vertical fluxes

ii) Inertial layer - in which turbulent fluid motions dominate the vertical fluxes, but the dominant scales of motion are still much less than the boundary layer depth. This is the layer in which most surface wind measurements are made.

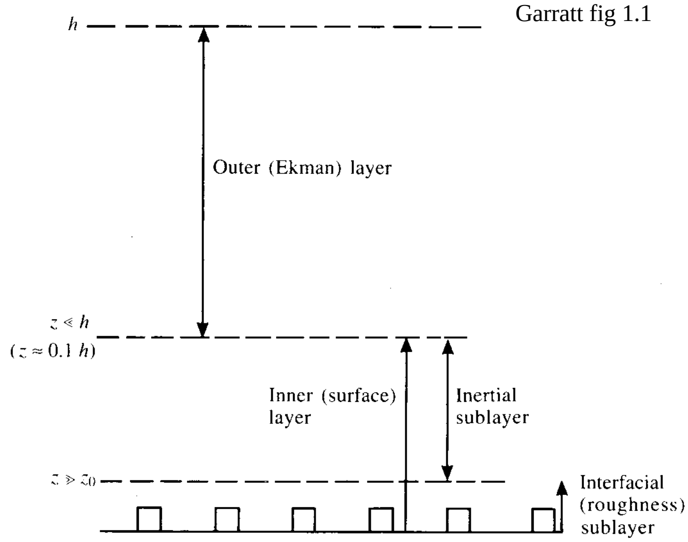
- Layers (i) + (ii) comprise the surface layer. Coriolis turning of the wind with height is not evident within the surface layer.

iii) Outer layer - turbulent fluid motions with scales of motion comparable to the boundary layer depth ('large eddies').

- At the top of the outer layer, the BL is often capped by an *entrainment zone* in which turbulent BL eddies are entraining non-turbulent free-atmospheric air. This entrainment zone is often associated with a stable layer or inversion.
- For boundary layers topped by shallow cumulus, the outer layer is subdivided further into

subcloud, transition, cumulus and inversion layer.

Garratt fig 1.1



Boundary layers are classified as *unstable* if the air moving upward in the turbulent motions tends to be buoyant (less dense) than in the downdrafts, and *stable* if the reverse is true. If there is negligible buoyancy transport within the BL, it is called *neutral*. On a hot sunny morning, surface heating causes the boundary layer to become strongly unstable, and convect vigorously with outer layer updrafts of 1-3 m s^{-1} which are a few tenths of a K warmer than the downdrafts, transporting several hundred W m^{-2} of heat upward. In desert regions such BLs can grow to a depth of 5 km or more by afternoon, though typical summer early afternoon BL depths over Midwest, Seattle, etc. are 1-2 km. At night, the surface cools by radiation. The BL depth can become as little as 50 m on a clear calm night, and the BL tends to be *stable*, with weak downward buoyancy fluxes. Rarely is an ideal neutral ABL observed, but with strong winds, buoyancy effects can become relatively unimportant, especially for winds over the oceans blowing along contours of constant SST.

Typical ABLs over the ocean tend to be slightly unstable, with little diurnal cycle due to the near-constancy of SST. BL depths vary from a few hundred m in regions of warm advection to 1.5-3 km where cold advection has led to shallow cumuli (subtropical trade wind belts, cold air outbreaks). In regions of deep convection, a BL top can be difficult to define.

Within the ocean, there is also an oceanic BL driven by surface wind stress and sometimes convection, and considerably affected by the absorption of radiation in the upper ocean. It is usually but not always stable. The oceanic BL can vary from a few m deep to a few km deep in isolated locations (e. g. Labrador Sea) and times where oceanic deep convection is driven by intense cold air advection overhead.

Applications and Relevance of BLM

The boundary layer is the part of the atmosphere in which we live and carry out most human activities. Furthermore, almost all exchange of heat, moisture, momentum, naturally occurring particles, aerosols, and gasses, and pollutants occurs through the BL. Specific applications

i) *Climate simulation and NWP* - parameterization of surface characteristics, air-surface exchange, BL thermodynamics fluxes and friction, and cloud. No climate model can succeed without some consideration of the boundary layer. In NWP models, a good boundary layer is critical to proper prediction of the diurnal cycle, of low-level winds and convergence, of effects of complex terrain, and of timing and location of convection. Coupling of atmospheric models to ocean, ice, land-surface models occurs through BL processes.

ii) *Air Pollution and Urban Meteorology* - Pollutant dispersal, interaction of BL with mesoscale circulations. Urban heat island effects.

iii) *Agricultural meteorology* - Prediction of frost, dew, evapotranspiration.

iv) *Aviation* - Prediction of fog formation and dissipation, dangerous wind-shear conditions.

v) *Remote Sensing* - Satellite-based measurements of surface winds, skin temperature, etc. involve the interaction of BL and surface, and must often be interpreted in light of a BL model to be useful for NWP.

History of BLM

1900 - 1910
- Development of laminar boundary layer theory for aerodynamics, starting with a seminal paper of Prandtl (1904).
- Ekman (1905,1906) develops his theory of laminar Ekman layer.

1910 - 1940
- Taylor develops basic methods for examining and understanding turbulent mixing
- Mixing length theory, eddy diffusivity - von Karman, Prandtl, Lettau

1940 - 1950
- Kolmogorov (1941) similarity theory of turbulence

1950 - 1960
- Buoyancy effects on surface layer (Monin and Obuhkov, 1954)
- Early field experiments (e. g. Great Plains Expt. of 1953) capable of accurate direct turbulent flux measurements

1960 - 1970
- The Golden Age of BLM. Accurate observations of a variety of boundary layer types, including convective, stable and trade-cumulus. Verification/calibration of surface similarity theory.

1970 - 1980
- Introduction of resolved 3D computer modelling of BL turbulence (large-eddy simulation or LES). Application of higher-order turbulence closure theory.

1980 - 1990
- Major field efforts in stratocumulus-topped boundary layers (FIRE, 1987) and land-surface, vegetation parameterization. Mesoscale modeling.

1990 - The Age of Technology
- New surface remote sensing tools (lidar, cloud radar) and extensive space-based coverage of surface characteristics;
- LES as a tool for improving parameterizations and bridging to observations.
- Coupled ocean-atmosphere-ice-biosphere and medium-range forecast models create stringent accuracy requirements for BL parameterizations.
- Accurate routine mesoscale modelling for urban air flow; coupling to air pollution
- Boundary layer - deep convection interactions (e. g. TOGA-COARE, 1992)

Why is the boundary layer turbulent?

We characterize the BL by turbulent motions, but we could imagine a laminar BL in which there is a smooth transition from the free-tropospheric wind speed to a no-slip condition against a surface (e .g. a laminar Ekman layer). Such a BL would have radically different characteristics than are observed.

<u>Steady Ekman BL equations</u> (z = height, surface at z = 0, free troposphere is z) :

$$-fv = \nu\, d^2u/dz^2$$

$$f(u - G) = \nu\, d^2v/dz^2$$

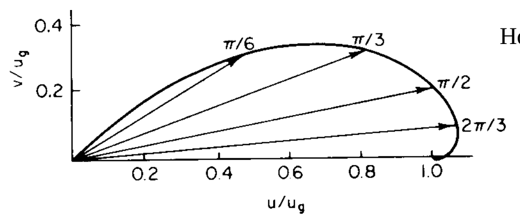
$$u(0) = 0,\ u(\infty) = G$$

$$v(0) = 0,\ v(\infty) = 0$$

Solution ($\zeta = z/\delta$) for BL velocity profile

$$u(z) = G(1 - e^{-\zeta} \cos \zeta)$$

$$v(z) = G\, e^{-\zeta} \sin \zeta$$



Holton, fig. 5.4

Fig. 5.4 Hodograph of the wind components in the Ekman spiral solution. The arrows show the velocity vectors for several levels in the Ekman layer, while the spiral curve traces out the velocity variation as a function of height. Points labeled on the spiral show the values of γz, which is a nondimensional measure of height.

Flow adjusts nearly to geostrophic within Ekman layer depth $\delta = (2\nu/f)^{1/2}$ of the surface.With a free tropospheric (geostrophic) velocity of G in the x direction, the kinematic molecular viscosity of air $\nu = 1.4\times10^{-5}$ m^2 s^{-1} and a Coriolis parameter $f = 10^{-4}$ s^{-1} , δ= 0.5 m, which is far thinner than observed!

Hydrodynamic Instability

Laminar BLs like the Ekman layer are not observed in the atmosphere because they are *hydro-dynamically unstable*, so even if we could artificially set such a BL up, perturbations would rapidly grow upon it and modify it toward a more realistic BL structure. Three forms of hydrodynamic instability are particularly relevant to BLs:

i) Shear instability

ii) Kelvin-Helmholtz instability

iii) Convective (Rayleigh-Benard) instability

By examining these types of instability, we can not only understand why laminar boundary layers are not observed, but also gain insight into some of the turbulent flow structures that are observed. The

<u>Shear Instability</u>

Instability of an unstratified shear flow $U(z)$ occuring at high Reynolds numbers Re = VL/ν, Tritton

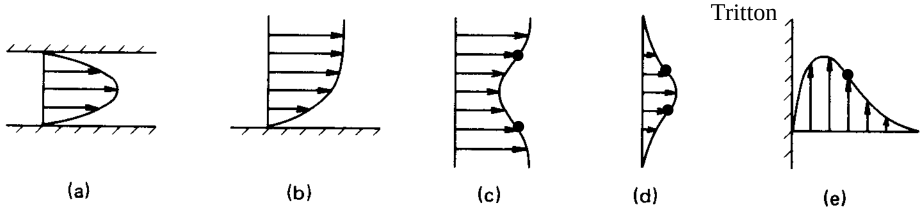


Figure 17.13 To illustrate that the velocity profiles of (a) pipe flow, (b) a boundary layer, (c) a wake, (d) a jet, and (e) a free convection boundary layer are all shear flows.

Some shear flows. Dots indicate inflection points.

where *V* is a characteristic variation in the velocity across the shear layer, which has a characteristic height *L*. Here, 'high' means at least 10^3; an ABL with a shear $V = 10$ m s^{-1} through a boundary layer of depth 1 km would have

$$\mathrm{Re} = (10 \text{ m s}^{-1})(1000 \text{ m})/ (10^{-5} \text{ m}^2 \text{ s}^{-1}) = 10^9,$$

which is plenty high!

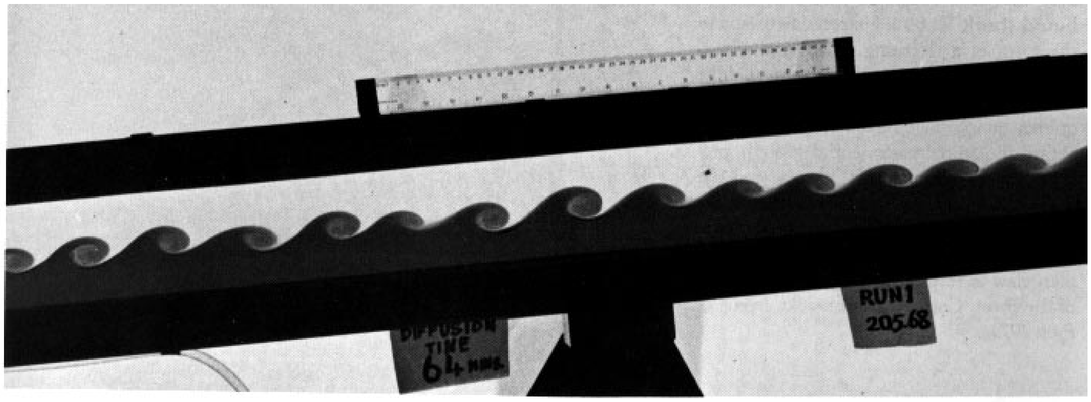
Inviscid shear flows can be linearly unstable only if they have an *inflection point* where $d^2U/dz^2 = 0$ (Rayleigh's criterion, 1880) and are definitely unstable if the vorticity dU/dz has an extremum somewhere inside the shear layer, not on a boundary (Fjortoft's criterion, 1950). This excludes profiles such as linear shear flows or pipe flows between boundaries, but some such profiles are in fact unstable at small but nonzero viscosity, and may still break down into turbulence. The Ekman layer profile has an inflection point, so is subject to shear instability (as well as a second class of instability at moderately large Re of a few hundred).

In shear instability a layer of high vorticity rolls up into isolated vortices. A good example is the von Karman vortex street that forms the the wake behind a moving obstacle.

94. Kármán vortex street behind a circular cylinder at R=140. Water is flowing at 1.4 cm/s past a cylinder of diameter 1 cm. Integrated streaklines are shown by electrolytic precipitation of a white colloidal smoke, illuminated by a sheet of light. The vortex sheet is seen to grow in width downstream for some diameters. *Photograph by Sadatoshi Taneda*

van Dyke, p. 56

Kelvin-Helmholtz Instability



145. Kelvin-Helmholtz instability of stratified shear flow. A long rectangular tube, initially horizontal, is filled with water above colored brine. The fluids are allowed to diffuse for about an hour, and the tube then quickly tilted six degrees, setting the fluids into motion. The brine accelerates uniformly down the slope, while the water above similarly accelerates up the slope. Sinusoidal instability of the interface occurs after a few seconds, and has here grown nonlinearly into regular spiral rolls. *Thorpe 1971*

van Dyke, p. 85

For an inviscid stratified shear layer with an inflection point, instability of the shear layer may still occur if the stratification is sufficiently weak. Shear instability at the interface between two layers of different densities was first investigated by Helmholtz (1868). Miles (1960) showed that for a continuously varying system, instability cannot occur if the static stability, as measured by buoyancy frequency N is large enough that

$$\mathrm{Ri} = N^2/(dU/dz)^2 > 1/4 \text{ throughout the shear layer}$$

For lesser values of Ri, instability usually does occur. The general form of this criterion can be rationalized by considering the mixing of two parcels of fluid of volume V at different heights. In a flow relative coordinate system:

Lower parcel has height $-\delta z$, initial density $\rho - \delta\rho$, velocity $-\delta U$.

Upper parcel has height δz, initial density $\rho + \delta\rho$, velocity δU.

Here $\delta U = (dU/dz)\delta z$, and $\delta\rho = (d\rho/dz)\delta z$, where $N^2 = -(g/\rho)(d\rho/dz)$. For simplicity we consider an incompressible fluid, and assume each parcel has volume V, at heights. The total initial energy of the parcels is

$$\begin{aligned} E_i &= KE_i + PE_i \\ &= 0.5V\{(\rho - \delta\rho)(-\delta U)^2 + (\rho + \delta\rho)(\delta U)^2\} + V\{(\rho - \delta\rho)g(-\delta z) + (\rho + \delta\rho)g(\delta z)\} \\ &= V\{\rho(\delta U)^2 + 2g\delta\rho\delta z\} \end{aligned}$$

If the parcels are homogenized in density and momentum,

Lower parcel has height $-\delta z$, final density ρ, velocity 0.

Upper parcel has height δz, final density ρ, velocity 0.

The total final energy is

$$\begin{aligned} E_f &= KE_f + PE_f \\ &= 0 + V\{\rho g(-\delta z) + \rho g(\delta z)\} = 0, \end{aligned}$$

so the change in total energy is:

$$\Delta E = E_f - E_i = -V\{\rho(\delta U)^2 + 2g\delta\rho\delta z\}$$
$$= V\rho(\delta z)^2\{-(dU/dz)^2 + 2N^2\}.$$

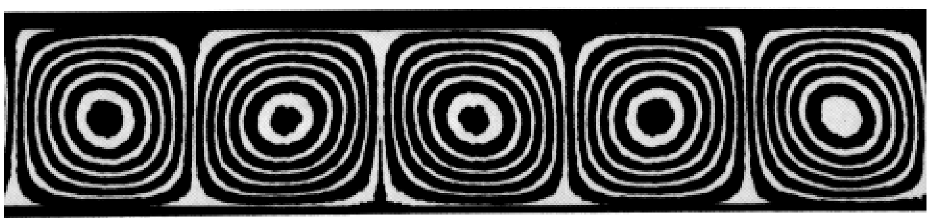
An energy reduction occurs if $(dU/dz)^2 > 2N^2$, i. e. if Ri < 1/2. In this case, residual energy is available to stir up eddy circulations. The reason this argument gives a less restrictive criterion for instability than an exact argument is that momentum is not fully homogenized in instabilities of a shear layer.

Convection

Thermal convection occurs if the potential density decreases with height in some layer. Classically, this instability has been studied by considering convection between two parallel plates in an incompressible fluid. The lower plate is heated to a fixed temperature that is larger than that of the upper plate. In the absence of convection, the temperature profile within the fluid would vary linearly with height due to conduction. If the plates are a distance h apart and have a temperature difference ΔT, and if the fluid has kinematic viscosity ν and thermal diffusivity κ ($= 2\times10^{-5}$ m^2 s^{-1} for air), then convective instability occurs when the Rayleigh number

$$\mathrm{Ra} = h^3\Delta B/\nu\kappa > 1700$$

Here ΔB is the buoyancy change $-g\Delta\rho/\rho$ associated with a temperature increase of ΔT at a given pressure; for air and other ideal gasses, $\Delta B = g\Delta T / T$. The instability is a circulation with cells with comparable width to height, a property of thermal convection observed even when Ra is much larger. Rolls and hexagonal patterns are equally unstable.



Slightly unstable convection in silicone oil van Dyke p. 82

In the presence of a mean shear, the fastest growing convective instabilities are rolls aligned along the shear vector, as seen in the cloud streets below.

Fig. 7.2. Convective clouds in an unstable layer, aligned in 'streets' along the direction of shear. (Compare with fig. 4.14 pl. x, which shows clouds formed by a shear instability and aligned across the flow. The form of 'billow' clouds can vary widely according to the relative importance of shear and convection.) (Photograph: R. S. Scorer.) Turner

For ABL convection, the surface skin temperature can be a few kelvins warmer than the typical boundary layer air temperature. Even with a small ΔT = 1 K, we can estimate ΔB = (10 m s^{-2})(1 K)/(300 K) = 0.03 m s^{-2}, h = 1000 m, and

$$\mathrm{Ra} = (0.03 \text{ m s}^{-2})(1000 \text{ m})^3 / (1.4 \times 10^{-5} \text{ m}^2 \text{ s}^{-1})(2 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}) = 10^{17} \; !$$

so the atmosphere is very far indeed from the instability threshold due to the large lengthscales and small viscosities.

Transition to turbulence

Each of these instabilities initially has a simple, regular circulation pattern. However, if the fluid is sufficiently inviscid, three-dimensional secondary instabilities grow on the initial circulation, and the flow becomes complex, irregular in time, and develops regions in which there are motions on a variety of scales. This is a transition into turbulent motion. We don’t generally see this transition in the ABL, since the ideal basic state on which the initial instability grows is rarely realized.

102. Instability of an axisymmetric jet. A laminar stream of air flows from a circular tube at Reynolds number 10,000 and is made visible by a smoke wire. The edge of the jet develops axisymmetric oscillations, rolls up into vortex rings, and then abruptly becomes turbulent. *Photograph by Robert Drubka and Hassan Nagib*