Lecture 17, Wind and Turbulence, Part 2, Surface Boundary Layer: Theory and Principles

Instructor: Dennis Baldocchi
Professor of Biometeorology
Ecosystem Science Division
Department of Environmental Science, Policy and Management
345 Hilgard Hall
University of California, Berkeley
Berkeley, CA 94720

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Topics to be Covered:

Variation in Space

1. Wind in the Surface Boundary Layer: Concepts and Principles
   a. Boundary Layers
      i. Planetary Boundary Layer
      ii. Surface Boundary Layer
      iii. Internal Boundary Layer (constant flux layer)
      iv. Flux footprints
      v. Nocturnal Boundary Layer
   b. Logarithmic Wind Profiles

L17.1 Variation in Space

L17.1.1 Boundary Layers

The earth's surface is the lowest **boundary** of the atmosphere. Transport processes associated with the transfer of heat, mass and momentum modify the properties of the lowest few kilometers of the atmosphere. A distinct aspect of the planetary boundary layer is its turbulent nature. In comparison, turbulence is rarely found in the upper atmosphere, except where shear occurs, as in the jet stream. Visual evidence of low shear and turbulence in the upper atmosphere is obtained via the persistence of jet contrails long after a jet has passed overhead.

Study of the planetary boundary layer has much importance and significance to the topic of biometeorology for numerous reasons.

1. it's where we live

2. fifty percent of the atmosphere's kinetic energy is dissipated in the boundary

3. it is the location of the source and sink of many trace gases, including water vapor, CO_2, ozone, methane

4. it is a reservoir of trace gases and pollutants

In classical fluid dynamics, boundary layers are described as the layer between the surface and the free stream flow where u is less than 99% of the free stream velocity. In micrometeorology and biometeorology, several boundary layers are defined.

L17.2.2. Planetary Boundary Layer

The planetary boundary layer is the layer of air near the ground that responds to spatial and temporal changes in the properties of the surface. This layer is turbulent and is well mixed. Its height evolves with time over the course of a day. Its maximum height can reach 3 km over deserts, dry fields and boreal forests. Over wetter surfaces the PBL reaches about 1 to 2 km. It consists of a **surface boundary layer**, a **well-mixed layer** and a **capping entrainment layer**. A conceptual view of the planetary boundary layer is shown below.

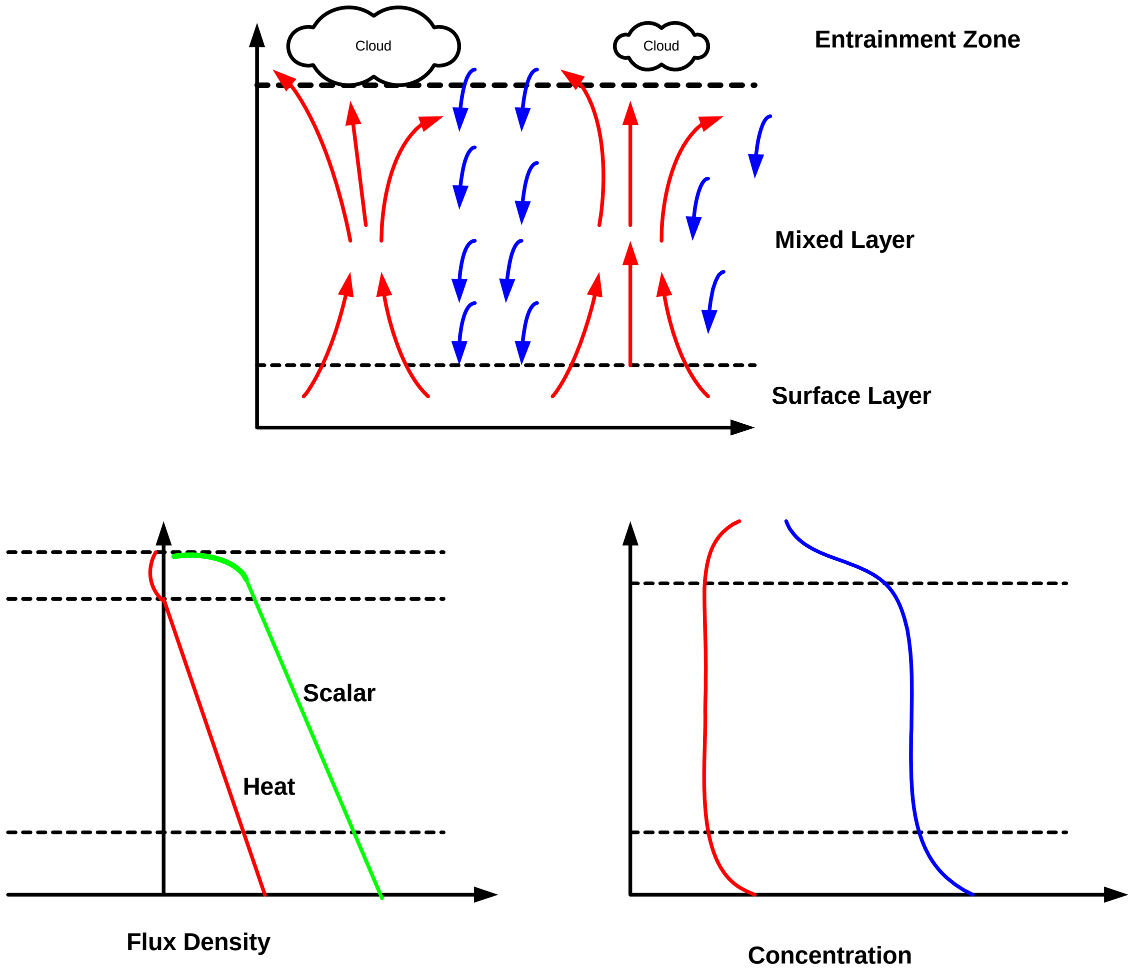


Figure 1 Conceptual view of the planetary boundary layer

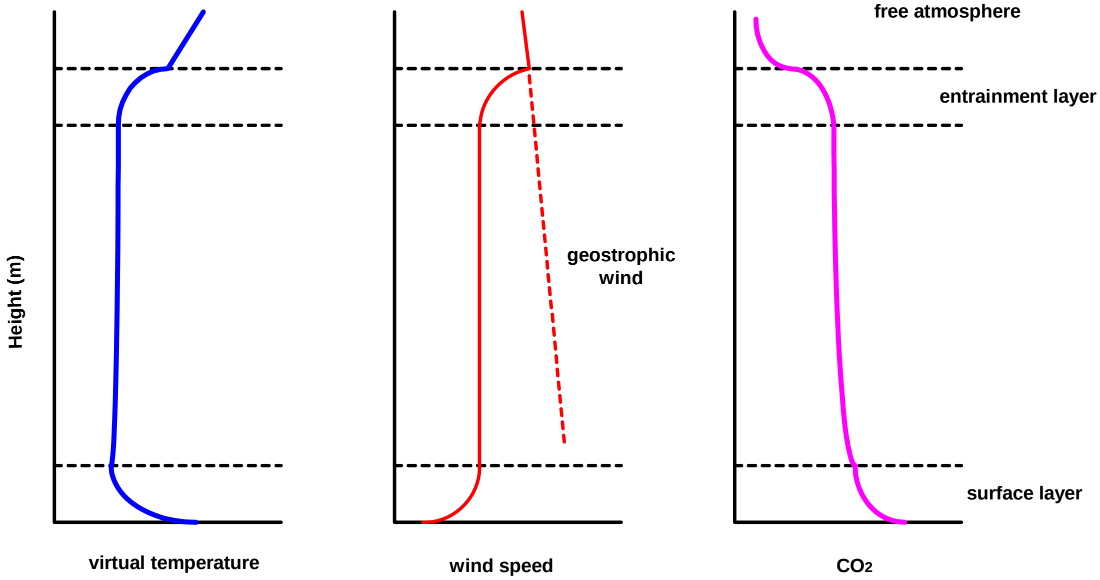


Figure 2 Conceptual profiles of temperature and wind in the surface, mixed and entrainment boundary layers. Adapted from Stull(Stull, 1988)

Also shown is data on wind and temperature profiles in the planetary boundary layer.

L17.2.3. Surface Boundary Layer

The **surface layer** is the lowest layer of the atmosphere (and troposphere). It is the layer, where air is in contact with the surface and where **strong vertical gradients** in temperature, humidity, wind and scalars exist. The atmosphere responds to surface forcing on the time scale of an hour or less. Temperature, wind, CO_2, humidity and pollutant concentrations exhibit distinct diurnal patterns over the course of a day in the troposphere. (The CO_2 variation relates to the rectifier effect and how CO_2 measurements are used in global inversion models to deduce sources and sinks.)

Wind direction does not change with height and the **Coriolis force** is nill. The depth of the surface boundary layer is about 10% of the planetary boundary layer.

L17.2.4. Internal Boundary Layer (constant flux layer)

The **internal boundary layer** is the layer of air in immediate contact with the surface. It is often called the **constant flux** layers, since the transfer of heat, momentum and mass is invariant with height. The development of the depth of the *ibl* is a function of the surface roughness and the distance from the edge. One algorithm for estimating its height is:

$$\delta(x) = 0.1x^{4/5}z_0^{1/5}$$

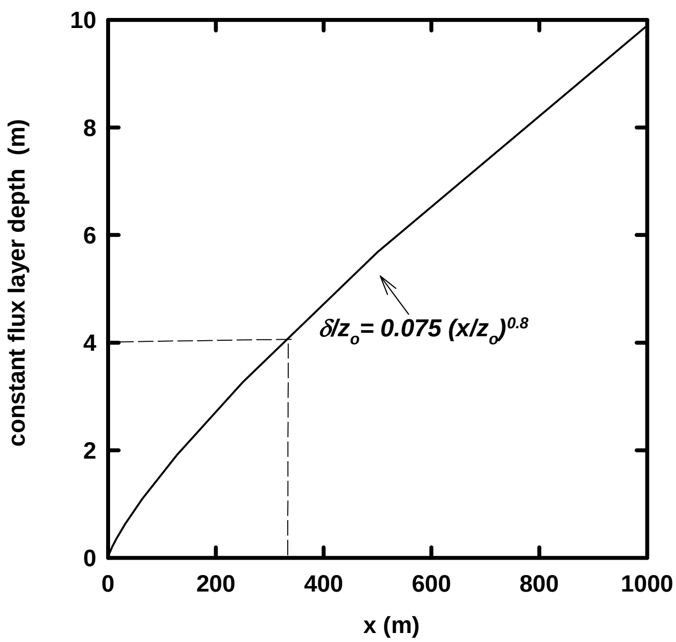


Figure 3 Computed evolution of the constant flux layer

L17.2.6 Nocturnal/stable Boundary Layer

The **nocturnal boundary** layer is ill behaved and difficult to quantify. Typically it is associated with low wind speeds near the surface as the air layer above has a thermal profile that is stably stratified. This effect, however, causes elevated jetting and sporadic turbulence and momentum from above must ultimately be absorbed by the surface ((Mahrt, 1999). Attributes of the stable boundary layer that must be quantified include radiative heat loss on clear nights, drainage of air with undulating terrain, shear instabilities and intermittent turbulence, generation of gravity waves, and the formation of fog and or dew.

L17.2.6 Wind profiles above vegetation

If we stand in a field on a windy day, we can feel strong wind forces on our face. But, when we reach down and touch the ground or lay on the grass, we feel that the wind speed is diminished. In fact, we notice that the wind speed goes to nearly zero at the ground surface. From this qualitative perspective, we can deduce that wind speed varies with height. But how wind speed vary quantitatively with height?

From dimensional arguments we can state that the shear is inversely proportional to height.

$$\frac{\partial u}{\partial z} = \frac{a}{z}$$

Because the vertical shear ($\frac{\partial u(z)}{\partial z}$) is inversely related to height, it is greatest near the ground and decreases with height.

Conceptually, **shear** is defined as the sum of translation, rigid rotation, and pure deformation (Blackadar, 1997).

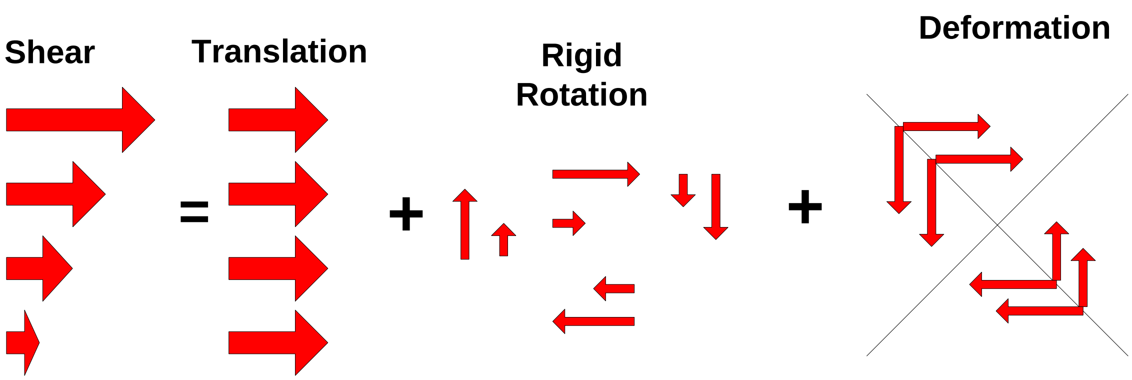


Figure 4 Conceptual diagram of shear and its contributors. Adapted from Blackadar, 1997.

How do we quantify the proportionality coefficient that defines how wind speed varies with height about a vegetated surface?

Given that wind speed will approach zero at the surface and the free stream velocity at some height above the surface, we have a shearing stress that is equal to the drag force per unit area that is exerted by the Earth and the plant canopy. The shear stress has units of mass times acceleration per unit area (N m^{-2}=kg m^{-1} s^{-2}).

In other words, a **force** is needed to change momentum transfer from one level to another. The drag force at the Earth's surface is one such example. This **drag force** or shear stress is also equivalent to the **momentum flux density** (mass times velocity per unit area per unit time (kg x m s^{-1} x m^{-2}).

One can envision momentum flux in a manner analogous to mass flux as computed with Fick's Law. If we have more momentum at a higher layer ($\rho u(z_2)$) than a lower layer ($\rho u(z_1)$), a potential gradient in momentum will be established. Momentum will flow from the region of **higher momentum** to that of **lower momentum**. From a dimensional viewpoint, the momentum flux is equal to a density (kg m^{-3}) times a velocity squared (m^2 s^{-2}). We can also use two extreme boundary points to infer that momentum will be transferred from the atmosphere to the ground surface. We can accept as given that the wind a certain distance above a surface is moving at a 'free-stream' velocity. We also know from common experience that somewhere near the ground the wind velocity is zero. So, momentum must be transferred downward.

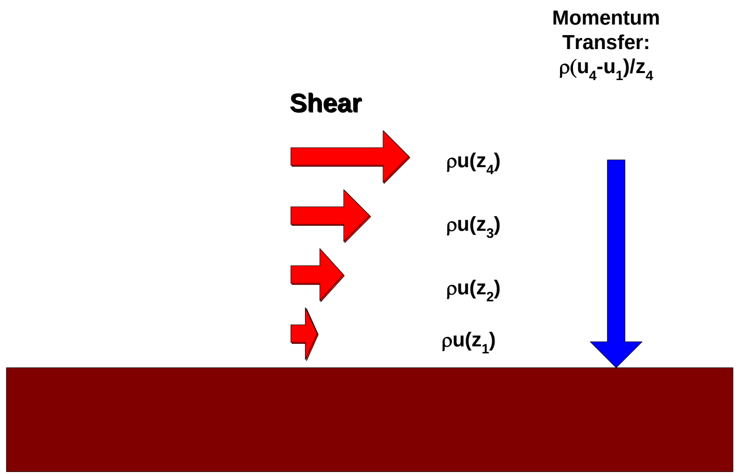


Figure 5 Visualization of momentum transfer

Thereby, from physical argument we can deduce that the **wind shear** is a function of **momentum transfer, height** above the surface and **air density**.

$$\frac{du}{dz} = f(\tau, \rho, z)$$

Greater shear stress will cause greater wind gradients, and vice versa. Density is related to momentum transfer, as it defines the mass per unit volume of air.

The Reynolds shear stress only occurs when the atmosphere is turbulent. A shear stress occurs because wind having different velocities is layered over one another. Hence the wind velocity gradient is proportional to the ratio of momentum transfer and height:

$$\frac{du}{dz} \propto \frac{\tau}{z} \propto \frac{u_*}{z}$$

We need to maintain proper units, hence we conveniently define and use the **friction velocity,** u_*. Subsequently, the wind shear is defined as:

$$\frac{du}{dz} = \frac{u_*}{kz}$$

The product kz is analogous to a mixing length, an indicator of the mean eddy size. The coefficient, k, is **von Karman's constant**. It has a value near 0.40 in most boundary layer flows.

In recent years the von Karman constant has been subject to periodic scrutiny over the years. Reviews (Foken, 2006; Hogstrom, 1988; Hogstrom, 1996) reports values of k between 0.35 and 0.42. The overall conclusion of over 18 studies is that *k* is constant, close to 0.40, and invariant of the friction Reynolds number between 0.1 and 10^5.

Another way to deduce the theoretical description of the wind profile gradient is to start with flux gradient theory and **Prandtl's mixing layer theory**. If a parcel of air with a mean horizontal velocity is moved to an upper layer with a velocity its wind fluctuation can be expressed by:

$$u' = \overline{u}(z-l) - \overline{u}(z) = -l\frac{\partial u}{\partial z}$$

Prandtl also assumed that w'~-u'

l is the mixing length, the distance eddies travel before they mix and lose their identity.

$$l \approx z$$
$$l = kz$$

Momentum transfer, defined as the w u covariance can be expressed as:

$$\overline{w'u'} \sim -\overline{l}^2 (\frac{\partial \overline{u}}{\partial z})^2$$

Adopting an analog to Fick's Law of Diffusion for momentum transfer we can define an eddy exchange coefficient for momentum transfer ($m^2\ s^{-1}$):

$$K_m = l^2 \frac{\partial \overline{u}}{\partial z} = k^2 z^2 \frac{\partial u}{\partial z} = u_* kz \quad (\mathrm{m}^2\ \mathrm{s}^{-1})$$

Re-arranging equations we have:

$$\tau = \rho \overline{w'u'} \approx \rho K_m \frac{\partial u}{\partial z}$$

Note that the Flux gradient equation is an approximation of the actual flux density of momentum. This is a first-order closure approximation for momentum transfer, a second order term with respect to turbulence.

$$(\frac{\partial u}{\partial z})^2 = \frac{\overline{w'u'}}{k^2 z^2}$$

The shear stress is often evaluated in terms of the friction velocity

$$\tau = \rho u_*^2$$

where **friction velocity** is defined as:

$$u_* = |\overline{w'u'}|^{1/2} \quad (\text{m s}^{-1})$$

where u is the longitudinal wind velocity. If one is using a three-dimensional anemometer, prior to rotating the horizontal coordinate system so u is aligned with X rather than x, then

$$u_* = \sqrt{(\overline{w'u'}^2 + \overline{w'v'}^2}$$

$$\frac{\partial u}{\partial z} = \frac{u_*}{kz}$$

Integrating the wind shear equation

$$\int_0^U \partial u = \frac{u_*}{k} \int_{z_0}^{Z} \frac{\partial z}{z}$$

yields an expression for wind speed, the famous **logarithmic wind profile** relation:

$$u(z) = \frac{u_*}{k} ln(\frac{z}{z_0})$$

It needs to be stressed at this stage that the wind velocity gradient and its logarithmic function are valid only for very short vegetation and for a neutrally stratified atmosphere.

The new variable, z_0, is the **roughness parameter**. It defines the effectiveness of a canopy to absorb momentum. Conceptually, it defines the height where the extrapolated wind flow approaches zero, and is the lower limit of the integrand. In practice, it approximates the height at which the fluid flow changes from being turbulent to being laminar. Typically, the roughness length is 10% of the height of the surface elements. Consequently, one cannot state categorically that z_0 is the height of the turbulent/laminar transition. There will be a zone of turbulent flow below the roughness height. We also stress that this relation is only valid for cases where $z >> z_0$.

If we know the wind speed on one height, we can compute it at other height by taking the ratios, thereby eliminating the need to assess the friction velocity

$$u(z_2) = u(z_1)\ln(\frac{z_2}{z_0}) / \ln(\frac{z_1}{z_0})$$

We can also use this equation and solve directly for z_0. It must be stressed here and at other points, these relations are only valid for near neutral stratification:

$$ln\, z_0 = \frac{u_2\, ln\, z_1 - u_1\, ln\, z_2}{(u_2 - u_1)}$$

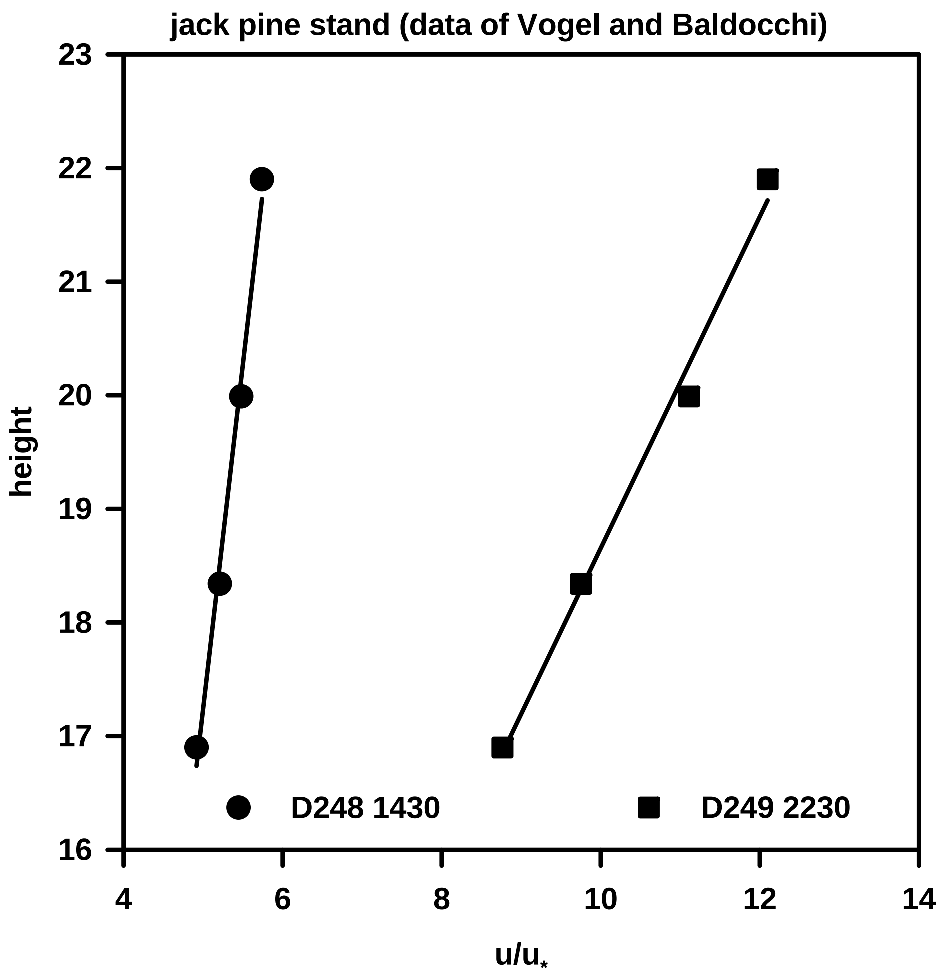
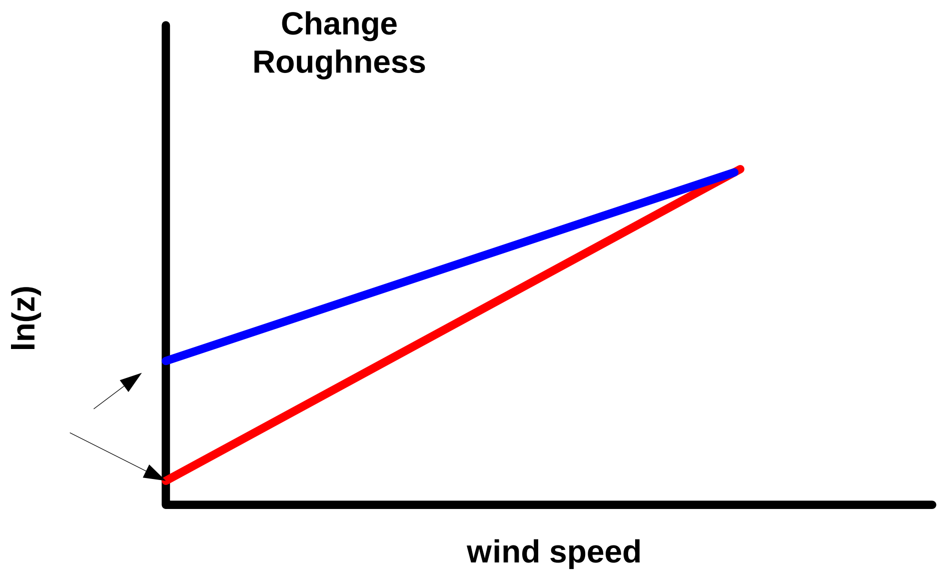


Figure 6 Wind profile measurements over a 13 m tall jack pine stand

Visually, a change in roughness increases the shear, du/dz, and increases the rate of momentum transfer to the surface.



An empirical understanding of scaling relations not only helps model turbulence, set boundary conditions, but it helps to debug measurements in the field. These relations tell you what to expect!!!

The drag coefficient is one such variable. It is defined from:

$$\tau = \rho C_d u^2$$

C_d increases with surface roughness. In many meteorological applications it is defined at 10 m. For a lake it is 0.0012. Note there are also definitions of the drag coefficient based on u/u*, the square root of the former definition. In this form, 0.1 to 0.2 is a typical value over vegetation.

Summary points

- Several boundary layers are pertinent to the study of atmospheric turbulence. They are the planetary boundary layer, the surface boundary layer and the internal or constant flux layer.
- A nocturnal boundary layer exists at night with its own distinct properties, due to the stability of the surface layer. It is associated with decoupling between the surface and upper layer, there can be a jetting of winds aloft and turbulent transfer can be intermittent and associated with gravity waves.
- Momentum transfer is related to the covariance between vertical velocity and horizontal velocity fluctuations. It can also be estimated using flux-gradient theory:

$$\tau = \rho \overline{w'u'} = \rho \cdot u_*^2 \approx \rho K_m \frac{\partial u}{\partial z}$$

- Friction velocity is a scaling velocity that is related to momentum transfer

$$u_* = \sqrt{\frac{\tau}{\rho}}$$

- The eddy exchange coefficient for momentum transfer can be estimated with measurements of wind profiles:

$$K_m = k^2 z^2 \frac{\partial u}{\partial z} = u_* kz \ \ (\mathrm{m}^2\ \mathrm{s}^{-1})$$

- Wind velocity profiles in the surface boundary layer is a logarithmic function of height. The slope of the log profile for wind velocity is a function of friction velocity and the zero intercept is a function of the roughness length (z_0); k is von Karman's constant (0.4):

$$u(z) = \frac{u_*}{k} \ln(z / z_0)$$

- For a similar wind velocity at height z (u(z)), friction velocity increases with increasing surface roughness, eg z_0.

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