GEOG 321 - Reading Package Lectures 19 & 20

KINETIC ENERGY

Horizontal temperature variations in the Earth-Atmopshere system give rise to horizontal pressure differences, which result in motion (winds). In this way thermal energy from the solar energy cycle is converted into *mean kinetic energy* of wind systems. The energy then participates in the energy cascade involving the transfer of energy to increasingly small scales of motion into *turbulent kinetic energy*. Kinetic energy can enter the cascade at a size-scale governed by the forces generating the motion. The energy is then passed down to smaller-sized eddies until it eventually is opposed by viscosity on a molecular scale and is dissipated as heat.

In the atmospheric boundary layer (ABL) we are concerned both with motion generated on the micro- and local scales, and with the modification of existing airflow generated on scales larger than those of the boundary layer. In the first category we are concerned with wind generated by horizontal thermal differences in the boundary layer. These local scale thermal winds are especially prevalent across the boundaries between contrasting surface types. Examples include the breezes occurring at land/sea (lake), mountain/ valley, forest/grassland, and urban/rural interfaces (see Lecture 30). In the second category we are concerned with the role of surface roughness in shaping the variation of wind speed with height, and with the way in which uneven terrain (e.g. hills and valleys) and isolated obstacles (e.g. a tree or a building) perturb existing flow patterns.

Most of our course is concerned with conditions close to Earth's surface where interactions between the atmosphere and biosphere are concentrated.

MOMENTUM

The momentum possessed by a body is given by the product of its mass and velocity. In the case of air the mean horizontal momentum of a *unit volume* is therefore given by its density (ρ) multiplied by its mean horizontal wind speed $\overline{u}$ (i.e., momentum of a unit volume = $\rho\,\overline{u}$). Since for practical purposes we may consider air density to be constant in the surface layer, the mean horizontal momentum possessed by different levels is proportional to the wind speed.

Drag and shear stress. The wind speed in the planetary boundary layer is largely controlled by the frictional drag imposed on the flow by the underlying rigid surface. The drag retards motion close to the ground and gives rise to a sharp decrease of mean horizontal wind speed ($\overline{u}$) as the surface is approached. The force exerted on the surface by the air being dragged over it is called the surface *shearing stress* (τ) and is expressed as a pressure (Pa, force per unit surface area). This force is equally opposed by that exerted by the surface on the atmosphere.

However since air is a fluid it only acts on the lower boundary and not throughout the total bulk of the atmosphere. The *surface layer* of frictional influence generates this shearing force and transmits it downwards as a *flux of momentum*.

In the absence of strong thermal effects the depth of the frictional influence depends on the roughness of the surface (Figure 1).

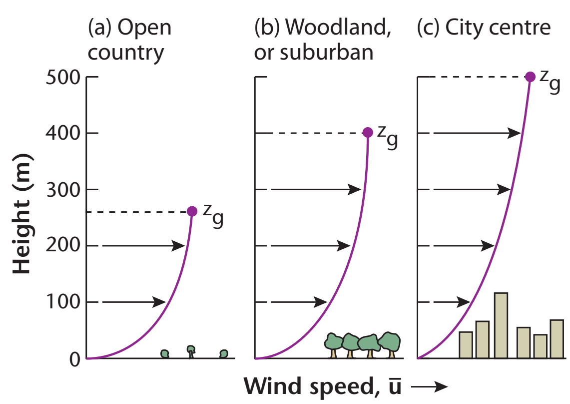

Figure 1: The wind speed profile near the ground illustrating the effect of terrain roughness (after Davenport, 1965).

The profiles in this figure are based on measurements in strong winds, and the height z_g is the top of the boundary layer above which $\overline{u}$ is approximately constant with height (i.e. surface drag is negligible). The depth of this layer increases with increasing roughness. Therefore the vertical gradient of mean wind speed ($\partial \overline{u} / \partial z$) is greatest over smooth terrain, and least over rough surfaces. In light winds the depth z_g also depends upon the amount of thermal convection generated at the surface. With strong surface heating z_g is greater than in Figure 1, and with surface cooling it is less.

Sweeps and ejections. Consider the situation at level z_3 in Figure 2b. Due to the effects of forced convection generated by the surface roughness, and the mutual shearing between air layers moving at different speeds, turbulent eddies are continually moving up and down through z_3. An eddy arriving at z_3 having originated at z_4 above will, upon mixing, impart a net increase in velocity (and hence momentum). A wind speed sensor at z_3 would therefore see this downdraft as an increase in wind speed, a 'gust' (called *sweep*). Conversely an updraft from z_2 would be sensed as an 'lull' (called *ejection*) in horizontal wind speed.

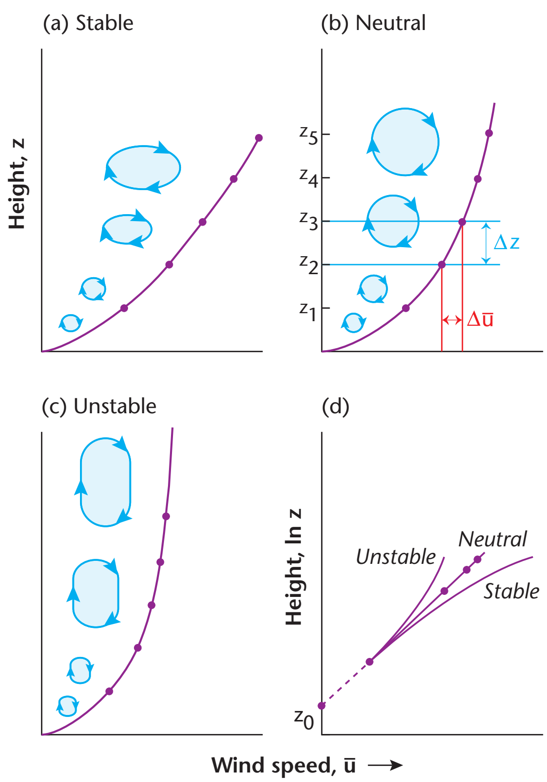

Figure 2: The effect of stability on the wind profile shape and eddy structure near the ground (after Thom, 1975). In (d) the profiles of (a) to (c) are re-plotted with a natural logarithm height scale.

Notice that due to the increase of wind with height the net effect of both updrafts and downdrafts is always to sustain a net flux of momentum downwards.

In terms of the general characteristics of the interaction between different surfaces and airflow we may expect that for a surface of given roughness τ will depend upon $\overline{u}$. As $\overline{u}$ at a reference level increases so will τ , and so will the depth of the forced convection layer. Similarly if we considered the value of $\overline{u}$ to be constant over surfaces with different roughnesses, the magnitude and depth of forced convective activity would be greatest over the roughest surface (Figure 1).

VERTICAL WIND PROFILE

The actual form of the wind variation with height under neutral stability has been found to be accurately described by a logarithmic decay curve. Thus using the natural logarithm of height ($\ln z$) as the vertical coordinate the data from Figure 2a fall upon a straight line in Figure 2d. This provides the basis for the logarithmic wind profile equation:

$$\overline{u}(z) = \frac{u_*}{k} \ln \left(\frac{z}{z_0} \right) \qquad \star \quad (8.1)$$

where $\overline{u}$ - mean wind speed ($\mathrm{m\,s^{-1}}$) at the height z, u_* - friction velocity ($\mathrm{m\,s^{-1}}$), k - von Karman's constant (≈ 0.4), z_0 - roughness length (m). It has been found that the shear stress τ is proportional to the square of the wind velocity at some arbitrary reference height. Thus we introduce for which this square law holds exactly so that:

$$u_*^2 = \frac{\tau}{\rho} \qquad \star \quad (8.2)$$

This is helpful because u_* can be evaluated from wind profile measurements (the slope of the line in Figure 2d is k/u_*) and therefore we can obtain τ, which can be used in evaluating other fluxes (see Lecture 24).

Roughness length. The length z_0 is a measure of the aerodynamic roughness of the surface. It is related, but not equal to, the height of the roughness elements. It is also a function of the shape and density distribution of the elements. Typical values of z_0 are listed in Table 1. This term is defined as the height at which the neutral wind profile extrapolates to a zero wind speed (Figure 2d).

Table 1: Roughness length z_0 of natural surfaces.

Surface	Remarks	z_0 Roughness length (m)
Water†	Still – open sea	$0{\cdot}1–10{\cdot}0 \times 10^{-5}$
Ice	Smooth	$0{\cdot}1 \times 10^{-4}$
Snow		$0{\cdot}5–10{\cdot}0 \times 10^{-4}$
Sand, desert		0·0003
Soils		0·001–0·01
Grass†	0·02–0·1 m	0·003–0·01
	0·25–1·0 m	0·04–0·10
Agricultural crops†		0·04–0·20
Orchards†		0·5–1·0
Forests†	Deciduous	1·0–6·0
	Coniferous	1·0–6·0

† z_0 depends on wind speed (see p.139)
Sources: Sutton (1953), Szeicz *et al.* (1969), Kraus (1972).

The effects of stability. The foregoing discussion relates to neutral conditions where buoyancy is unimportant. Such conditions are found with cloudy skies and strong winds, and in the lowest 1 to 2 m of the atmosphere. Cloud reduces radiative heating and cooling of the surface; strong winds promote mixing and do not permit strong temperature stratification to develop; and in the lowest layers forced convection due to frictionally-generated eddies is dominant. In the simplest interpretation these eddies may be conceived as being circular and to increase in diameter with height (Figure 2b). In reality they are three-dimensional and comprise a wide variety of sizes.

In unstable conditions the vertical movement of eddies (and therefore the momentum flux) is enhanced. Near the surface mechanical effects continue to dominate but at greater heights thermal effects become increasingly more important. This results in a progressive vertical stretching of the eddies and a reduction of the wind gradient (Figure 2a). Conversely strong stability dampens vertical movement, progressively compresses the eddies and steepens the wind gradient (Figure 2c).

Stability effects on turbulence are further illustrated in Figure 3. This is a graph of wind inclination (roughly corresponding to vertical winds, because the scale refers to the tilt angle of a horizontal vane) over a period of 3 minutes. The upper trace is from lapse (unstable), and the lower trace from inversion (stable) conditions, over the same grass site with approximately equivalent horizontal wind speeds. Thus differences between the two traces are due to stability differences.

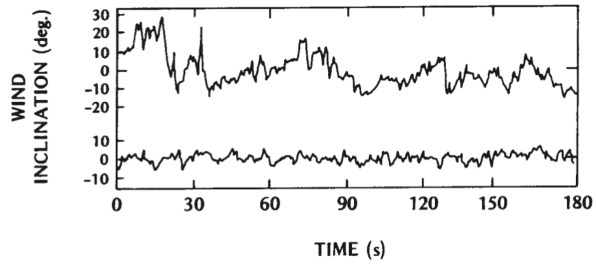

Figure 3: The effect of stability on the turbulent structure of the atmosphere. Wind inclination fluctuations at a height of 29 m during unstable (upper trace) and stable (lower) conditions over a grass site with winds of 3 to 4 m s^{-1} (after Priestley, 1959).

In the unstable case two types of fluctuation are evident. First, there are long-period 'waves' lasting about 1 to 1.5 minutes. These are relatively large buoyancy-generated eddies bursting up through the measurement level (positive values) or being replaced by sinking air parcels (negative values). Superimposed on this pattern are a second set of much shorter-period fluctuations. These are the small roughness-generated and internal shearing eddies. Therefore Figure 3 visually presents the two elements of turbulence free convection (large) and forced convection (small). The combination of these two elements (upper trace) provides a very efficient means of both vertical transport and mixing. This is the 'ideal' daytime mixed convection situation. The stable case (lower trace) by contrast only exhibits the short-period eddies due to forced convection because buoyancy is absent. This is the 'ideal' nocturnal situation, and is not conducive to vertical exchange.

In summary we may say that below approximately 2 m the effects of forced convection dominate even in non-neutral conditions as long as there is a reasonable airflow. Above this height the relative role of free convection grows and the possibility of stability effects on momentum transfer increases. These effects are manifested as curvature in the wind profile (Figure 2a-c). Strong instability weakens the wind gradient by promoting vertical exchange over a deep layer, and thereby mixing the greater momentum of faster-moving upper air with that nearer the surface. Strong stability on the other hand strengthens the wind gradient. It therefore follows that since there is a characteristic diurnal cycle of stability there is an associated diurnal variation of wind speed in the surface layer.

Zero plane displacement. A plot of wind speeds measured at a number of levels above tall vegetation results in a profile such as that in Figure 4. Above the vegetation the wind profile is logarithmic as with other surfaces, but the extrapolation of this curve downwards shows that the flow is behaving as though the 'surface' is located at some height near the top of the stand, and not at the ground. This height is called the level of *zero plane displacement* (d), which may be visualized as representing the apparent level of the bulk drag exerted by the vegetation on the air (or the level of the apparent momentum sink).

In practice for a wide range of crops and trees the value of d is approximately given by:

$$d \approx \frac{2}{3}h \tag{8.3}$$

where h is the height of the stand. Equation 8.4 only applies to closely-spaced stands because d also depends on the density of the drag elements. The value of d also depends upon the wind speed because most vegetation is flexible and assumes a more streamlined form at high wind speeds. The logarithmic wind profile equation (equation 8.1) should therefore be modified over tall roughness to read:

$$\overline{u}(z) = \frac{u_*}{z} \ln\left(\frac{z-d}{z_0}\right) \quad \star \tag{8.4}$$

The study of momentum exchange and the form of the wind profile is important because of what it tells us about the state of turbulence. This is central to questions concerning the transport of heat and water vapour and to the dispersal of air pollutants.

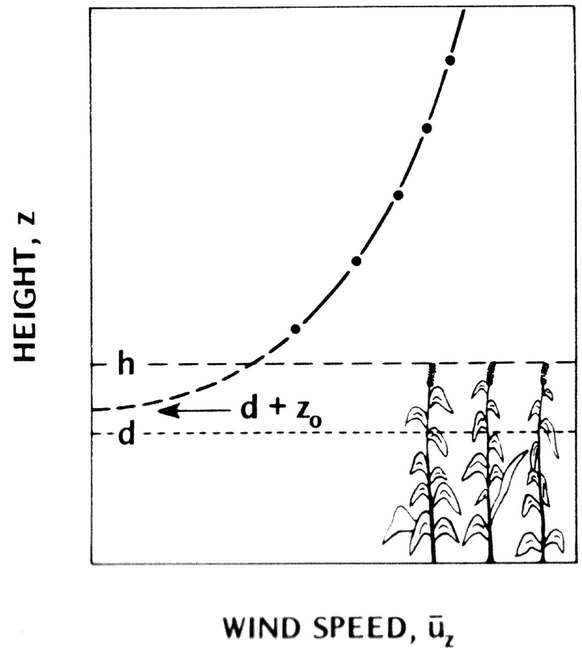

Figure 4: Typical wind profile measured above a vegetation stand of height h, illustrating the concept of a zero plane displacement at the height d.