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Surface Layer

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Introduction

The lowest layer of the atmosphere, the surface layer, directly influences the daily activities of nearly all life on planet Earth. Extending up to an altitude of order 50–100 meters, it possesses physical and chemical properties long recognized as having a controlling influence over a wide range of human and societal interests. For example, in 600 BC, the force of the wind acting against the flow of the Nile was already being used as a secondary process included in flood forecasts in ancient Egypt. By 2000 years ago, simple climatologies of near-surface wind directions were the basis of weather forecasts in the Mediterranean, for use in both military and commercial activities. By the late Middle Ages in both Europe and China, sailing ships were being designed to exploit the greater wind speeds which regularly were observed at increasing heights above the surface. In fact, the Chinese used kites to study surface layer wind profiles as early as 1400 years ago. More recently, i.e., in the early twentieth century, pollutant dispersion models were designed, based on anthropogenic emissions from point sources (such as smokestacks) and wind statistics, in order to estimate human exposure and deposition downwind. Some of the greatest challenges we face today in areas such as remote sensing involve understanding the relationship between the surface layer and the underlying terrestrial biosphere and oceans.

The tight interplay of the atmospheric surface layer and its involvement in surface energy exchanges is an essential ingredient in understanding both local and global climatology. The driving local process is a large downward flux of solar radiation, absorbed at the earth's surface during the day, and which in turn is converted rapidly to heat. While some of the solar heating is transferred within and down into the surface canopy, the highly turbulent nature of the surface layer over (low-heat-capacity) terrestrial surfaces allows a rapid upward flux of heat back into the atmosphere. Because the upward flux of heat is due to turbulence rather than to the less efficient laminar diffusion process, it averts the huge daily temperature extremes which would otherwise be encountered (such as on the Moon). On the other hand, during clear nights, the upward flux of infrared radiation combined with nearly no turbulence during calm conditions often leads to the more extreme cold nights. Note that if the nocturnal surface layer is otherwise windy, the associated higher turbulence levels would provide an efficient mechanism for transferring heat rapidly down to the surface, thus reducing the rate of nighttime cooling. From both a scientific and a social perspective, it is the presence or absence of turbulence in the surface layer which acts as a control over the degree to which humans and surface-based ecosystems are exposed to diurnal and climatic extremes.

In the atmospheric sciences, the depth of the surface layer is, by convention, defined as being 10% of the depth of the full planetary boundary layer (PBL) (**Figure 1**). With a PBL depth typically in the range of several hundred to 1500 m deep and capped by an overlying inversion, the wind speed, temperature, and humidity change rapidly with height throughout the surface layer; they then approach more constant values with height until they reach the inversion. The wind direction, in contrast, is relatively constant with height in the surface layer, then begins to change direction rapidly as height increases. The change of wind direction with height is due mostly to the

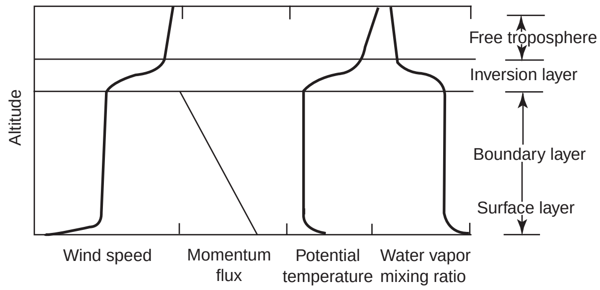

Figure 1 Depiction of profiles of windspeed, momentum flux, temperature and water vapor in the idealized planetary boundary layer, including surface layer.

influence of the Earth's rotation. As one might expect, owing to the generally higher wind shear near the surface, the turbulent downward momentum flux is maximum at the surface and approaches zero with height until it reaches the top of the PBL. While the surface layer can thus be characterized by vertical gradients of, e.g., wind speed and temperature, it has no identifiable top.

The surface layer may also be defined to be the lowest layer where the momentum flux is within 90% of its surface values. For this reason, the surface layer becomes a 'nearly' constant flux layer. Other flux gradients, such as those for temperature and gases, may behave differently, e.g., owing to thermodynamic influences or chemical reactions. However, because the vertical gradients of windspeed, temperature, and gas concentrations tend to be orders of magnitude larger than vertical gradients of their counterpart fluxes, many practitioners treat the surface layer as equivalent to a 'constant flux' layer. The reader is reminded that this is only an approximation.

In the next section, a brief survey of surface layer turbulence is given. This is followed by a description of similarity theory and its applications, sampling methods, and a survey of the theory of trace gas exchange. The summary will highlight recent and new applications of surface layer research.

Surface Layer Turbulence

While wind speed increases with height (owing to surface frictional effects), the surface layer is characterized by shear induced turbulence, which acts to transfer momentum, heat, and mass vertically within the air column. The strength of the turbulence, however, depends on other characteristics, namely surface roughness and atmospheric stability. For example, for a given wind speed over a rough surface, e.g., where forests or houses are present, the vertical mixing of momentum and heat is rapid and efficient compared with over smoother surfaces such as the ocean. The surface roughness also has a controlling influence on the shape of the wind and temperature profiles, which in turn affects the magnitude of surface fluxes.

Atmospheric stability plays a quite different role. Unstable stratification, such as one often finds on a hot summer afternoon, augments the turbulence levels with extra energy provided by the buoyant thermal plumes and eddies caused by surface heating from the sun. At night, the thermal eddies are generally absent, thus leading to stratified conditions and suppressed levels of turbulence. In the simplest form of surface layer theory, e.g., for neutral stratifications, turbulent eddies can be best described as simple wind velocity fluctuations with respect to the mean flow, where time and length scales of the fluctuations are associated with the local wind shear and height above the surface.

A time series of turbulent fluctuations will show that a simple view is insufficient to describe the real complexity of the surface layer. Unlike the simplistic view of shear creating characteristic eddies, field measurements have shown that eddies of many scales coexist at any given height. Scientists have therefore computed turbulence spectra, based on sampling height and characteristic surface layer properties, e.g., stability and roughness. Each of the eddies in the spectrum can, in turn, be assigned a kinetic energy. For those smallest-scale eddies (of order 1 mm), turbulent kinetic energy is quickly converted to heat, thus implying that the source of energy occurs at larger-scales. It therefore follows that energy flows from the largest-scale eddies to the smallest ones, through an 'energy cascade'. In practice, the eddies smaller than around a few meters are often assumed to be advected by the mean wind, yet also scale with height above the surface. Usually referred to as 'frozen' turbulence during the advection process, Taylor's hypothesis is often invoked for these small eddies, where the eddy's wavenumber, k_λ, can be related to its frequency, f, in Hz. Observed by fixed measurement sensors, the Taylor's hypothesis is expressed by the following relation:

$$k_\lambda = \frac{2\pi f}{U} \qquad [1]$$

where U is the mean windspeed.

The most important part of the turbulence spectrum for use in a wide variety of applications is the inertial subrange, which in turn includes turbulence scales spanning from order 1 cm up to around 1 m. Within this subrange, turbulent eddies also intimately participate in transferring energy to higher frequencies (to smaller scales), and one may assume that the spectral density, $S(k_\lambda)$, can be scaled to the rate of turbulent energy dissipation, ε, associated with the smallest-scale eddies. The classical form of the inertial subrange is written as

$$S(k_\lambda) = \alpha\varepsilon^{2/3}k_\lambda^{-5/3} \quad [2]$$

where $\alpha = 0.52$. The lowest frequency applicable to the scaling in eqn [2] is both height- and wind-speed-dependent, and a practical lower limit, in Hz, is U/z. In stark contrast to the small-scale eddies, the largest-scale eddies, of order 100 m or more, observed within the surface layer have sizes generally controlled by convective activity and other organized motions within the full planetary boundary layer. These large-scale eddies are sometimes even influenced by gravity waves within the inversion above the PBL.

The turbulence levels and fluxes within the surface layer may be written mathematically in terms of physical balances derived from the Navier–Stokes equation and other mass budget equations. The most important equation used for practical application to surface layer phenomena is the turbulent kinetic energy (TKE) budget, written as

$$\frac{\partial \bar{e}}{\partial t} = -U\frac{\partial \bar{e}}{\partial x} - \overline{u'w'}\frac{\partial U}{\partial z} - \frac{\partial \overline{w'e'}}{\partial z} - \frac{1}{\rho}\frac{\partial \overline{w'p'}}{\partial z} + \frac{g}{T_\mathrm{v}}\frac{\partial \overline{w'T'_\mathrm{v}}}{\partial z} - \upsilon\frac{\overline{\partial u'_i \partial u'_i}}{\partial x_j \partial x_j} \quad [3]$$

In [3], the quantities u', v', and w' represent the fluctuating downwind (x-direction), crosswind (y-direction), and vertical (z-direction) wind velocity components. $\bar{e}$ is TKE per unit volume (E), normalized by density (ρ), i.e.

$$\bar{e} = \frac{E}{\rho} = \frac{1}{2}\overline{u'_i u'_i} \quad [4]$$

and the instantaneous contribution to the TKE may be denoted as

$$e' \equiv \frac{1}{2}u'_i u'_i \quad [5]$$

The virtual temperature, T_v, is defined as $T(1 + 0.61q)$, where T is temperature and q is humidity. The first term on the r.h.s. of [3] represents the rate of change of TKE due to advection. The second term represents shear production. The third term is the flux divergence of TKE, while the following term is the divergence of pressure flux. The fifth term on the r.h.s. is TKE production or loss due to buoyancy, and the last term is the loss of TKE due to viscosity.

Because the complexity of eqn [3] prohibits easy application, simplifications are introduced to reduce the set of independent variables so that solutions may readily be obtained. The most notable simplifications are the assumptions of steady state, horizontal homogeneity of all quantities, mean vertical velocity of zero, and lateral symmetry in turbulence and fluxes with respect to the mean downwind direction. This leads to:

$$-\overline{u'w'}\frac{\partial U}{\partial z} = \frac{g}{T_\mathrm{v}}\frac{\partial \overline{w'T'_\mathrm{v}}}{\partial z} - \varepsilon + \bar{I} \quad [6]$$

where the last term on the r.h.s. of [6], i.e., the imbalance, is the sum of the terms representing the pressure flux divergence and the energy flux. While most experimentalists assume that the imbalance term can be neglected during most conditions, there is a body of research in recent years which suggests that the imbalance term is often large and must not be ignored. To date, there have been no parameterizations of $\bar{I}$ published in the literature for easy application.

The solution to eqn [6], where $\bar{I}$ is unfortunately assumed to be negligible in nearly all cases, has provided applications in acoustic and radar scattering associated with turbulence, and this equation serves as the basis for measurements of fluxes via the 'dissipation technique' on towers, ships, offshore buoys, and aircraft. The simplification of ignoring $\bar{I}$, however, has introduced substantial uncertainty in the use of eqn [6] in some regions of high scientific interest, e.g., in coastal zones and in heterogeous terrain.

Similarity Theory and Bulk Aerodynamic Relations

For many of the more important applications of surface layer theory, e.g., in estimating loads on structures and in support of pollutant dispersion calculations, information on the vertical profiles of wind speed, temperature, humidity, refractivity, and turbulence levels is necessary. In addition, estimates of the surface fluxes based on easily obtained observations of mean surface layer quantities (such as wind speed, temperature, etc.) are highly desirable for both the surface and remote sensing communities as well as climate modeling. It was partly to satisfy these needs that similarity theory was developed.

Similarity theory is based on the assumption that dimensionless groups of variables may be arranged in terms of functional relationships to the flow field, and where the number of variables is reduced to a closed set for easy application. The surface layer has been best described by the similarity theory proposed by Monin and Obukhov, hereinafter referred to as MOS theory. In MOS theory, it is postulated that the fundamental scales required to represent processes within the surface layer are the friction velocity $u*$ (denoted as the square root of $\overline{-u'w'}$); the surface heat flux $\overline{w'T'_v}$; the height above the surface, z; and the buoyancy parameter, g/T_v. It was furthermore assumed that the surface layer is a constant flux layer, for momentum, heat, and gases.

Based on the original MOS theory presented in the early 1950s, the profiles of any bulk quantity, X, which satisfies the assumptions behind MOS theory (including a constant flux layer for each compound) may be written in the following form:

$$\partial X/\partial z = \frac{X^*}{kz}\phi_x \qquad [7]$$

where $X = (U, T_v, q, \text{etc.})$. The quantity X^* represents the turbulent vertical flux of X normalized by u^*, and the function ϕ_x depends on stratification, as will be described shortly. For windspeed and temperature, eqn [7] may be rewritten as

$$\frac{\partial U}{\partial z} = \frac{u^*}{kz}\phi_U \qquad [8a]$$

$$\frac{\partial T}{\partial z} = \frac{T^*}{kz}\phi_T \qquad [8b]$$

where the von Karman constant, k, has a value of 0.4. In eqns [8a] and [8b], the functions, ϕ_U and ϕ_T are formulated in terms of a dimensionless stability parameter, z/L, where z is measurement height and L is the Monin–Obukhov length. The dimensionless stability parameter is based on the ratio of the shear-induced TKE to the buoyancy-induced TKE, i.e., it is proportional to the ratio of the first term on the r.h.s. of eqn [6] to the term on the l.h.s. of eqn [6]. Thus one may write

$$z/L = \frac{gkz\overline{w'T'_v}}{T_v u^{*3}} \qquad [9]$$

Integration of eqn [8a] will yield a logarithmic profile of wind speed, if one specifies a value of zero for the wind speed at a small height above the surface denoted as z_0, i.e., the roughness length. For temperature, the integration of eqn [8b] requires setting the surface temperature at an analogous height above the surface, i.e., z_{0T}. It is generally assumed that the lower boundary 'roughness lengths' are related to the physical characteristics of the surface.

As an illustration, upon integration of eqns [8a, b] one obtains

$$U - U_0 = \frac{u^*}{k \ln z/z_0 - \Psi_m} \qquad [10a]$$

$$T - T_0 = \frac{T^*}{k \ln z/z_{0T} - \Psi_T} \qquad [10b]$$

where Ψ_m, and Ψ_T are stability functions, and $T^* = \overline{w'T'}/u^*$. For neutral stratifications, i.e., where $\partial T_v/\partial z = 0$, the stability functions equal zero; positive/negative values of Ψ_m and Ψ_T correspond to unstable/stable stratifications. The functional relationships between Ψ_m and ϕ_u, and Ψ_T and ϕ_T, may be found in most references following this article. Rearrangement of eqns [10a] and [10b] yields the bulk aerodynamic relations for momentum and temperature fluxes, e.g.,

$$u^{*2} = C_D(U - U_0)^2 \qquad [11a]$$

$$u^*T^* = C_H(U - U_0)(T - T_0) \qquad [11b]$$

where the quantities C_D and C_H are respectively the drag coefficient and the Stanton number, i.e.,

$$C_D = \left(\frac{k}{\ln z/z_0 - \Psi_m}\right)^2 \qquad [12a]$$

$$C_H = \left(\frac{k}{\ln z/z_0 - \Psi_m}\right)\left(\frac{k}{\ln z/z_{0T} - \Psi_T}\right) \qquad [12b]$$

A similar relation can be derived for the humidity profile and its coefficient, the Dalton number.

The beauty of the derivation of the drag and other flux coefficients is that they, at least in theory, do not depend on wind speed and are weakly dependent on height, roughness, and typical ranges of atmospheric stability. We note herein that strongly stable stratifications do, in fact, lead to much smaller fluxes, i.e., when compared with neutral conditions.

The flux coefficients are based on field observations, collected from platforms which are preferably stable and with minimal flow distortion. In the scientific literature, the coefficients are often reported under the condition of neutral stratification and with a standard height (usually 10 m above the surface) so that they may be compared easily with other data sets and/or integrated into models as general parameterizations.

Applications of MOS Theory to Sampling of the Surface Layer

Most modern research into the characteristics of the surface layer has relied on very complicated measurements. Flow distortion, weathering, platform motions, and the simplifying assumptions associated with specific methods have made both the measurements and subsequent interpretations of data difficult. Until the 1970s, the most common method of gathering data was via vertical profiles of the average wind speed, temperature, and humidity. This method relied on the rearrangement of eqn [10] into a form where (x, y) plots of windspeed versus the natural logarithm of height yielded information on the friction velocity and surface roughness simultaneously. As an example, for wind speed, the slope is (u^*/k) and the bias is $[(u^*/k)(\ln z_0 + \Psi_m)]$ when one plots data in a form where $(x, y) = (U - U_0, \ln z)$.

During the 1970s and 1980s, the sonic anemometer and other fast-response instruments for measuring temperature and humidity rapidly replaced the profile technique with directly measured fluxes using eddy correlation. This technique relies on the Reynolds averaging of the equations of motion, where an ensemble average and fluctuation about the average are substituted for each constituent variable. The turbulent flux $\langle w'x' \rangle$ is expressed in terms of both its mean and fluctuating parts, yielding a general form,

$$\mathrm{Flux} = \langle w \rangle \langle x \rangle + \langle w'x' \rangle \qquad [13]$$

where the vertical velocity $w = \langle w \rangle + w'$; and the parameter undergoing turbulent exchange is represented as $x = \langle x \rangle + x'$. By applying the constraint that $\langle w \rangle$ is zero, the turbulent fluxes reduce to their more common form,

$$\mathrm{Flux} = \langle w'x' \rangle \qquad [14]$$

For momentum, eqn [14] must be treated in its vector form and account for both the downwind and crosswind velocity components of the surface stress vector, $\boldsymbol{\tau}$:

$$\boldsymbol{\tau}/\rho = -\langle u'w' \rangle i - \langle v'w' \rangle j \qquad [15]$$

The constraint imposed on eqns [14] and [15] requires that the averaging time is long enough for $\langle w \rangle$ to approach zero, and for enough important atmospheric eddies to have been sampled to reduce the standard error of variance to an acceptable level. It is generally recommended that one averages the turbulence time series over periods of 30 min to 1 h. For heat, the density flux is written in analogy to eqn [14], as proportional to $\langle w'T_v' \rangle$, and water vapor flux is simply written as $\langle w'q' \rangle$.

During the 1980s and 1990s, the dissipation method became a popular sampling technique, owing to its relative insensitivity to platform motions (such as those of buoys) and shorter averaging time than for the eddy correlation technique. The dissipation technique is based on rapid sampling of the one-dimensional turbulence spectrum of constitutent 'x' and application of the turbulent budget equation applicable to 'x'. For the case of momentum fluxes, the TKE budget denoted in eqn [6] and the form of the inertial subrange denoted by eqn [2] must both be combined into a form where u^* is related to the level of energy densities in the inertial subrange of measured TKE spectra. Noting that $u^{*^3} = \varepsilon kz$, i.e., where the imbalance term in eqn [6] is ignored, one obtains

$$S(k_\lambda) = \alpha(kz)^{-2/3} u^{*^2} k_\lambda^{-5/3} \qquad [16]$$

In addition to momentum, the dissipation method has been applied to infer temperature, water vapor, and carbon dioxide fluxes. As techniques begin to emerge for other compounds where sampling frequencies are greater than 1 Hz, the dissipation method will no doubt gain popularity across a broad range of species and applications. The challenge, however, remains how to deal with imbalance terms, not only in the TKE budget but also in variance budgets for temperature and gaseous compounds.

Trace Gas Exchange

Simple models for the surface exchange of trace gases exhibit more complexity than the bulk aerodynamic relations described in the previous sections. Not only does turbulent transfer in the surface layer play a key role, but one must consider the reactions of the compound with both other atmospheric species and with the surface. For chemical compounds which have reaction time scales which are long compared with surface layer turbulence time scales, the constant flux layer assumption may be invoked. This applies to gases such as CO_2, O_2, and SO_2. However, if the reactions are relatively fast, in particular with respect to the time scales of surface layer turbulence, a flux divergence of the particular compound will occur. Species such as HNO_3 and NH_3 are examples, and the following equation must be solved in this case:

$$U\frac{\partial c}{\partial x} + \frac{\partial \langle w'c' \rangle}{\partial z} = S \qquad [17]$$

In eqn [17], S represents a production or loss of chemical concentration, c, based on chemical reactions. Because of the risk of other processes which are more important than turbulent transport, this balance tests the limits of similarity theory applicable to the surface layer. In spite of this shortcoming, the scientific community has proceeded to produce a body of literature in which most assumptions are ignored if one can use reference heights which are closer to the surface, i.e., where characteristic turbulence time scales are much smaller.

An alternative approach to measure gas fluxes is based on the concept of the deposition velocity, v_d. Here one uses various additive resistances associated with the flow of compound, c.

$$v_d = \langle w'c' \rangle / (c - c_s) = (r_a + r_b)^{-1} \qquad [18]$$

where r_a is the aerodynamic resistance governing turbulent transport of species c; and r_b is the surface resistance, governing the diffusion transport over the laminar sublayer. The surface concentration, c_s, is set to zero for many gases, exceptions being some of the biogenic related gases (e.g., CO_2 and NH_3) which can exhibit high surface concentrations in some media. In essence, eqn [18] implies that the turbulent transport associated with r_a represents the maximum possible deposition velocity for any species undergoing air–surface exchange ($= 1/C_D U$), and r_b is a correction factor which depends on the properties of the particular compound undergoing air–surface exchange.

The surface resistance, r_b, is more difficult to describe. The quantity r_b is based on the assumptions (1) that there exists a relatively homogeneous laminar sublayer at the surface, and (2) that the physical characteristics of the surface and the biological characteristics govern the rate of diffusion.

Flux Profile Relations for Quasi-inhomogeneous Conditions

The MOS theory is generally supported by a set of assumptions which requires averaging over spatial scales of order 25 km or more. In recent years, however, computer power has allowed much finer-resolution models to be developed, and there has been a challenge to surface layer theorists to relax the requirements of horizontal homogeneity, steady state, and long averaging, yet still produce accurate flux estimates on scales as small as 1 km. Several modified theories have been introduced, some of which employ results of large eddy simulations while others explore the use of 'flux footprints' and local internal boundary layers.

In regions which exhibit systematic horizontal variability as continuous functions, the flux profile relations may be written in a modified form. For wind speed, one may write

$$\partial U/\partial z = (u^*/kz)(\phi_u - R - S + W + G) \qquad [19]$$

where $R = \gamma z/z_0\, \partial z_0/\partial x$, $S = \beta\gamma(z/L)^2\, \partial L/\partial x$, $W = (kzU^2/2u^{*^3})\, \partial U/\partial x$, and $G = kzfV_g U/2u^{*^3}$. The quantity γ represents the local slope of vertically diffusing properties from upwind surface flux footprints. The term W_m has been shown to be substantially more important on scales less than 1 km over the coastal ocean. Over land, both may be competing strongly in terrain with variable roughness characteristics.

The concept of the flux footprint emerged during the last decade, and it has strongly influenced the requirement of local horizontal homogeneity. The flux footprint is an upwind elliptical surface patch which has roughness characteristics affecting the downwind turbulence, flux, and bulk profiles at some point above the surface, such as on a fixed mast. For neutral stratifications, the flux footprint affecting turbulence and fluxes is located upwind at a distance between 40 h and 100 h, where h is the height above the surface such as on a mast. For practical and simple applications, the slope of the vertically diffusing properties of the upwind surface footprint to the local measurement height is around $\frac{1}{70}$, for neutral stratifications.

Challenges and New Directions

The state of our present knowledge of the surface layer has been built from the similarity paradigm introduced in 1954 by Monin and Obukhov. This theory proved quite successful, particularly on spatial scales which are greater than 25 km and temporal scales of order 1 h. During recent years, however, the scientific challenges given by mesoscale and boundary modelers and other applied customers involve scales up to three orders of magnitude smaller than scales which the research community explored even two decades ago. Obviously, the problems of determining the momentum flux to swell and wind waves, and the problem of estimating ammonia deposition patterns near the edge of sensitive forested ecosystems, are good examples where the basic assumptions of similarity theory have become routinely violated. For this reason, one may expect that studies of surface layer theory will soon enter a new era, where advanced modeling techniques will be combined with multiscale *in-situ* and remote sensing systems.

See also

Air–Sea Interaction: Momentum, Heat and Vapor Fluxes. **Boundary Layers:** Neutrally Stratified Boundary Layer; Observational Techniques *In Situ*. **Land–Atmosphere Interactions:** Trace Gas Exchange. **Parameterization of Physical Processes:** Turbulence and Mixing.

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BOW ECHOS AND DERECHO

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Introduction

Windstorms produced by complexes of convective storms (thunderstorms) pose a significant hazard to life and property in many places of the world, especially during the spring and summer months. The largest and most long-lived of these events have been given a generic name of 'derecho', a term that originated in the late 1800s to refer to convective systems producing wide and long swaths of straight-line wind damage. Detailed studies of convective wind events, however, have shown that a vast majority are associated with a particular type of organized convective system, more popularily referred to as a 'bow echo'. This chapter describes the basic structures and environments associated with bow echoes and derechoes, and further highlights some of the recent research that clarifies the mechanisms critical to their development and maintenance.

Bow echoes and derechoes form a subset under the more general heading of mesoscale convective systems, which include squall lines, mesoscale convective complexes, and the like. In all of these cases, the system is envisioned to be composed of a sequence of relatively independent convective cells that contribute collectively to a larger system-scale structure. The individual convective cells can be ordinary cells, multicells, or supercells, as described elsewhere in the encyclopedia (*see* **Convective Storms:** Overview). In the following, we emphasize the system-scale attributes that have led to these particular systems being identified for their unique form of mesoconvective organization.

Bow Echoes

Bow echoes, originally referred to as a line echo wave pattern (LEWP), are most readily identified by a persistent bow-shape on a radar screen, and have become especially associated with the production of long, narrow swaths of damaging surface winds. Much of what we know observationally concerning bow echoes originated with Dr T. T. Fujita, who spent much of his career trying to characterize and understand the types of convective systems most apt to produce severe weather such as downbursts and microbursts.

A typical evolution and morphology of radar echoes associated with a severe bow echo is presented in **Figure 1**. The system usually begins as a strong

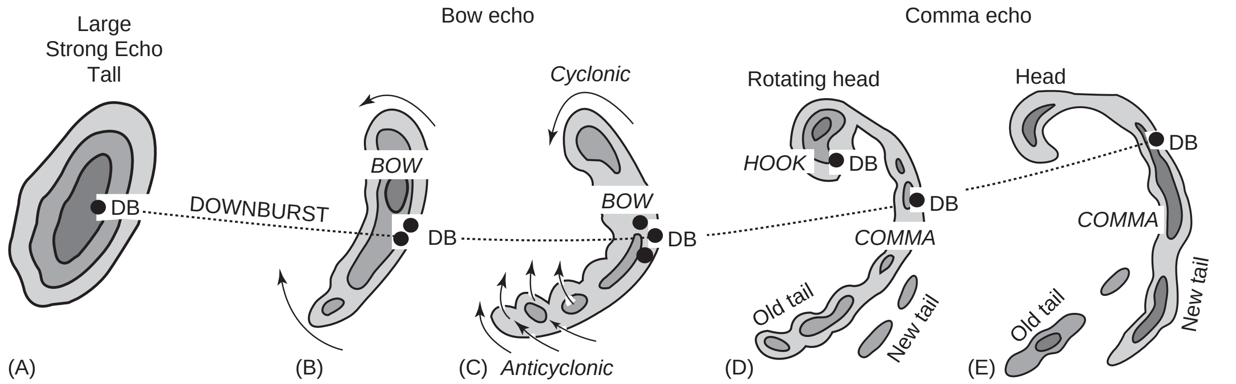

Figure 1 A typical morphology of radar echoes associated with bow echoes that produce strong and extensive downbursts, labeled DB on the figure. (Reproduced with permission from Fujita, 1978.)