A SURVEY OF OBSERVATIONS OF THE BOUNDARY LAYER

Roland B. Stull
Dept. of Atmospheric & Oceanic Sciences
University of Wisconsin, Madison, Wisconsin, USA

Abstract: Atmospheric boundary-layer (ABL) depth is constrained by a capping inversion, and does not scale to the Ekman depth. Coupling between the ABL and free troposphere are episodic – occurring during storms. Static stability and turbulence must be determined nonlocally. In fair weather over land, the ABL has a diurnal cycle with a deepening mixed layer during day, but at night it has a shallower stable boundary layer under a residual layer. Convective thermals cause anisotropic turbulence and entrainment into the top of the daytime mixed layer. The nocturnal stable boundary layer has weaker, sporadic turbulence, and can decouple from the residual layer above. Scaling variables are used to describe controls on the ABL, such as surface fluxes and ABL depth. Major research thrusts are currently extending our knowledge to more complex situations.

1. STRUCTURE

The atmospheric boundary layer (ABL) or planetary boundary layer (PBL) is the bottom 100 - 3000 m of the troposphere (Fig 1). The ABL is often turbulent and rapidly responds to surface forcings such as friction and other surface fluxes. The thickness of the boundary layer is quite variable in space and time, responding to the diurnal cycle of solar heating as well as to larger-scale dynamics and forcings.

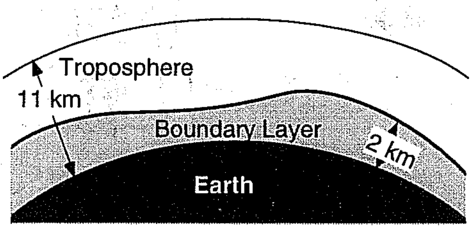

Fig. 1. Boundary layer location. (Stull, 1994)

A strong stable layer or temperature inversion usually marks the top of the ABL, and causes the remainder of the troposphere, loosely called the free atmosphere (FA), to be turbulently decoupled from the boundary layer. The reason is that the troposphere is statically stable on the average ($\partial\theta/\partial z \approx 3.3$ K/km.). Hence, turbulent mixing of the bottom of the troposphere creates a mixed layer (ML) that is cooler (in a potential temperature sense) than the rest of the troposphere, with a temperature inversion at the ABL top (Fig 2c). This inversion suppresses turbulence within it and traps most ABL turbulence below it, thereby causing dramatic decoupling from the free troposphere.

Paradigm Shift – ABL Depth

The stably-stratified troposphere provides a thermodynamic constraint on the ABL depth (Fig 2c) that dominates other depth scales. Engineering flows in pipes, channels, wind tunnels and over airfoils exhibit boundary layers that grow without bound (Fig 2a). On a hypothetical rotating earth with unstratified atmosphere, the balance between pressure-gradient, Coriolis and friction forces would create a theoretical depth scale called the Ekman-layer depth (Fig 2b): $h_E \propto u_*/f_c$, where u_* is the friction velocity and f_c is the Coriolis parameter. However, this Ekman depth u_*/f_c is **not** relevant to the real ABL.

The thermodynamic depth z_i of the ABL usually cannot be diagnosed, but must be forecast based on the heat budget and turbulence kinetic-energy considerations. This depth applies to ABLs of all static stabilities, even neutral stability.

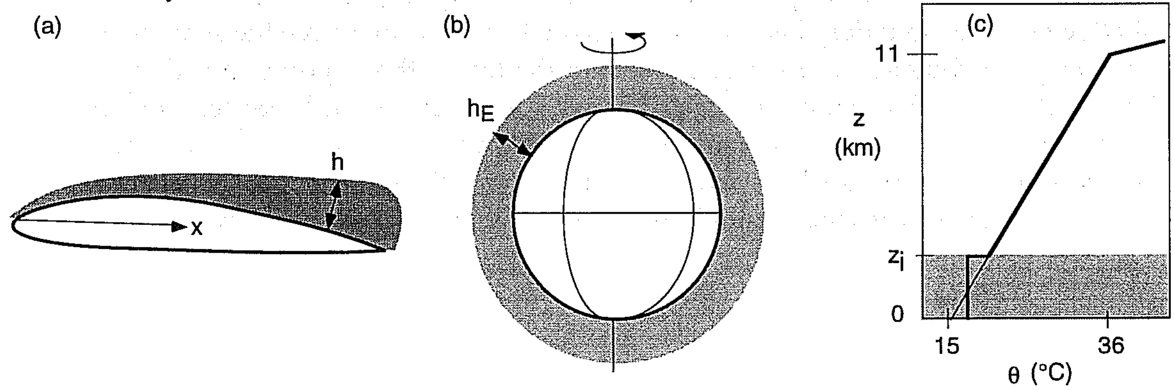

Fig. 2 Boundary layers (shaded) for (a) an engineering application, (b) a rotating earth, and (c) the troposphere.

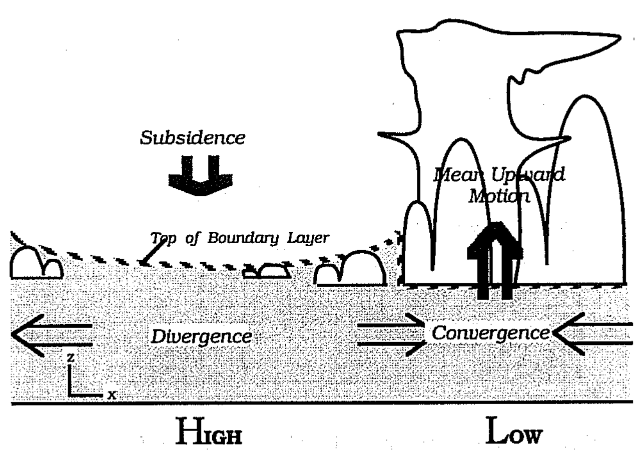

Fig. 3. Synoptic-scale modulation of the boundary layer.

The decoupling is most evident in anticyclonic conditions of large-scale subsidence (Fig 3). A very sharp top of the intense turbulence, high moisture, adiabatic lapse rate, and high pollution concentration marks the ABL top. Small forced clouds might be trapped within the ABL.

The opposite occurs in regions of bad weather and cyclones, where large-scale upward motion and surface convergence cause ABL air to rise in clouds to the top of the troposphere. This cloudy motion can be as convective thunderstorms, or as stratiform clouds where ABL air is pealed away from the surface and advectively driven up frontal surfaces.

Coupling between the ABL and FA is very episodic, patchy, and nonlinear. A reasonable idealization of the ABL is a reservoir within which trapped air interacts with the ground and is usually decoupled from the free atmosphere. However, localized vents (Fig 4) or pipes or conveyer belts occasionally but effectively remove ABL air and deposit it higher in the troposphere. Thus, the free atmosphere feels a free-slip, zero-flux bottom boundary.

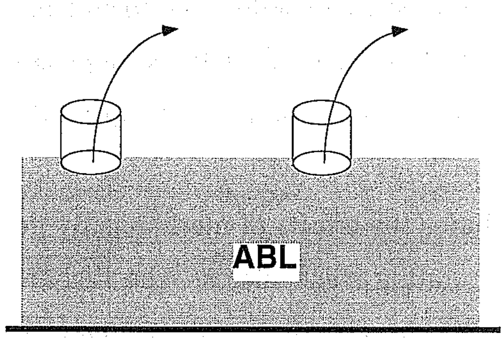

Fig. 4. Episodic interaction between PBL and remainder of the troposphere.

Replacement of air into the ABL from the FA happens via turbulent entrainment across the ML top, and via direct insertion or replacement of the whole ABL by storms. The entrainment process is controlled by the turbulence within the ABL, which erodes at the overlying air and incorporate sit into the ABL. As a result, the ABL increases in mass, volume, and depth. This is a one-way process into the top of the ML; as a result, pollutants and moisture already within the ABL are trapped. Clouds and storms are the exception. Deeper convective clouds can vent ABL air into the FA. Also, storms and fronts (Fig 5) can peal away ABL air and replace it with air from higher in the troposphere. Falling precipitation can also modify the ABL.

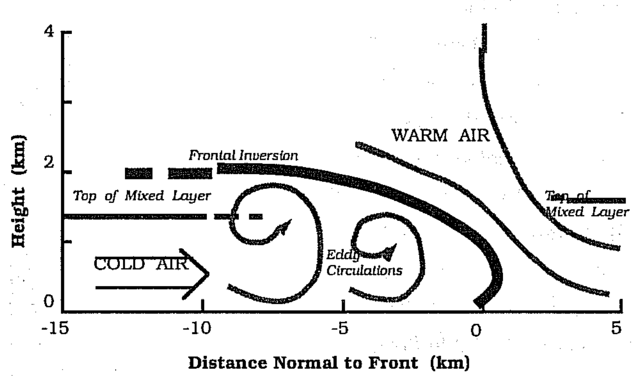

Fig. 5. Idealization of the boundary layer near a cold front.

2. STABILITY

Turbulence in the ABL is controlled by the flow stability. By definition, unstable flow becomes or remains turbulent, while stable flow becomes or remains laminar. In regions away from storms, the two most important measures of stability for the ABL are static and dynamic stability.

Turbulence usually causes mixing in such a way as to alleviate or undo the instability that caused it. Turbulence is also dissipative, meaning that turbulence will decay once the instability is eliminated. Thus, persistent turbulence exists in the atmosphere only where there is continued destabilization of the air, such as by solar heating of the ground or dynamic forcings that create wind shear.

Dynamic stability includes the effects of mechanically-generated turbulence along with buoyant (statical stability) generating mechanisms. The Richardson number is the usual local definition of static stability, with values less than about 0.25 implying turbulence.

The easiest way to avoid confusion in determining flow stability is that when different stability parameters give different stabilities, "unstable" always wins. The layer is unstable if either it is dynamically or nonlocally-statically unstable. The variation of static stability during a typical diurnal cycle causes variations in turbulent mixing, which allows different parts of the ABL to be identified.

Paradigm Shift – Static Stability

Most traditional definitions of static stability are incorrect, as they define static stability in terms of a local lapse rate. Within ABLs can be large coherent structures such as convective thermals. These eddies cause advective-like, not diffusive-like, turbulent transport, resulting in fluxes that go counter to the local gradient.

Static stability should be determined by conceptually lifting an air parcel (Fig 6) adiabatically from each relative maximum in a sounding, and conceptually lowering an air parcel adiabatically from each relative minimum. The vertical domain traveled by these air parcels until they hit another part of the sounding or the ground defines the domain of statically unstable air. Any local lapse rates within these unstable regions do not affect the static stability.

Only outside of the unstable regions can the local lapse rate be used to define static stability. This is done in the usual way with adiabatic profiles giving neutral stability, and subadiabatic profiles giving stable flow.

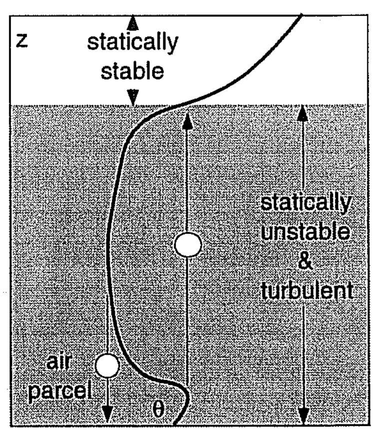

Fig. 6. Static stability determination.

3. STRUCTURE AND DIURNAL EVOLUTION

Over land in anticyclonic conditions with light winds, the diurnal cycle of solar heating of the ground creates a strong diurnal variation of the ABL (Fig 7). During daytime, a convective mixed layer (ML) grows with three stages: (1) early morning slow rise while the nighttime-chilled air is being heated and entrained; (2) rapid rise during late morning or early afternoon through the residual layer; and (3) deep ML of nearly constant depth (1 to 3 km) in the afternoon, where large-scale subsidence approximately balances entrainment (Stull, 1988). Turbulence is produced thermally by convection (thermals) and mechanically by wind shear. The temperature inversion at the top of the ML, through which entrainment occurs, is called the entrainment zone (EZ).

Near sunset when longwave radiative cooling of the surface commences. Thermal convection ceases and turbulence decays over much of the former ML. This layer is now called the residual layer (RL), and the now-nonturbulent stable layer is called the capping inversion (CI). The bottom of the residual layer is gradually converted into a stable boundary layer (SBL) during the night as shear-generated turbulence transfers heat from the air to the radiatively-cooled ground. The cycle then repeats during the next days until a storm comes. In real atmospheres, advection is always important and can even dominate. Typical soundings are shown in Fig 8.

3.1. Mixed layer and entrainment zone

Mixed layers (ML) were observed in the upper ocean, and the concept was applied to the atmosphere by Ball (1960). Deardorff's (1972, 1974, 1985 with Willis) pioneering work refined our understanding, and forms the core of the modern ML paradigm.

There is nearly constant virtual potential temperature(Hibbard & Sawford, 1994) over most of the ML interior (Fig 8). Mixing between the warm, dry, clean free atmosphere and the cooler, moist, polluted ML creates intermittent turbulence that is visible to remote sensors.

Conditional sampling by aircraft show that convective thermals dominate, with thermal diameters roughly equal to the mixed-layer depth, z_i (Fig 10). The mean state of updrafts and downdrafts differ, causing them to contribute differently to the net flux (Williams and Hacker, 1992). Other analysis methods include Fourier spectral analysis to focus on spacing between coherent structures, and wavelet analysis to focus on the shape of structures.

Turbulence is highly anisotropic, with most of the turbulence energy in the vertical. Temperature (and density) fluctuations are smallest in the interior of the ML . Fluxes such as heat flux generally vary linearly with height (Fig 9), as required to preserve the shape of the mean variable (potential temperature) within the mixed layer.

On a very local scale, the top of the ML is contorted as individual thermals punch up through the temperature inversion before sinking back into the ML. Averaged over many thermals, the top is surprisingly sharp and well defined. This average top may or may not follow valley and mountain topography, depending on the strength of the convection and bulk buoyancy in the mixed layer.

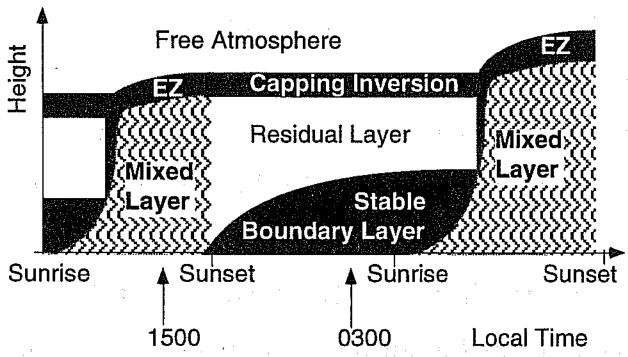

Fig. 7. (a) Idealized PBL evolution (Stull, 1994).

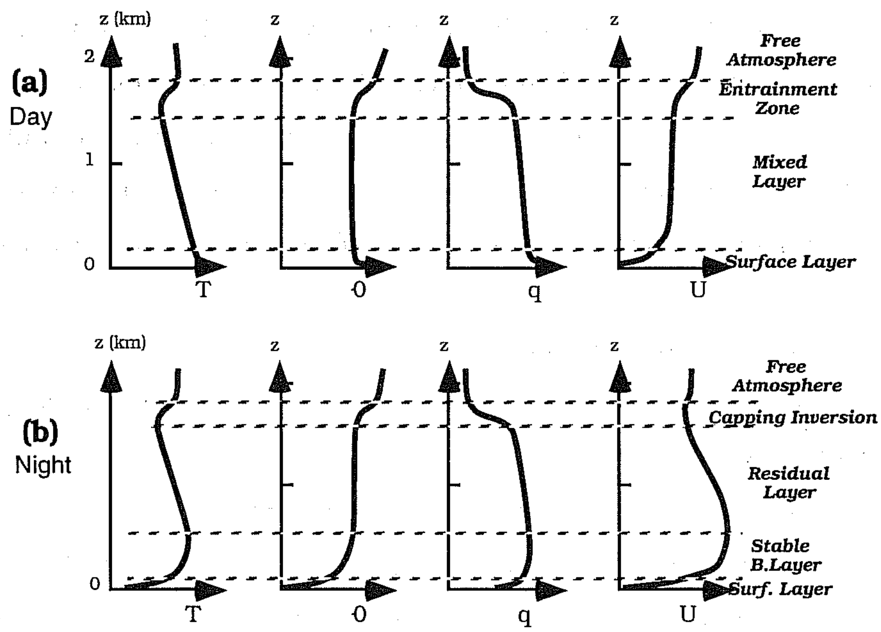

Fig. 8. (a) Idealized PBL evolution (Stull, 1994). (a) Daytime about 1500 local time. (b) Nighttime about 0300 local time. T = temperature, θ = potential temperature, q = mixing ratio, U = wind speed.

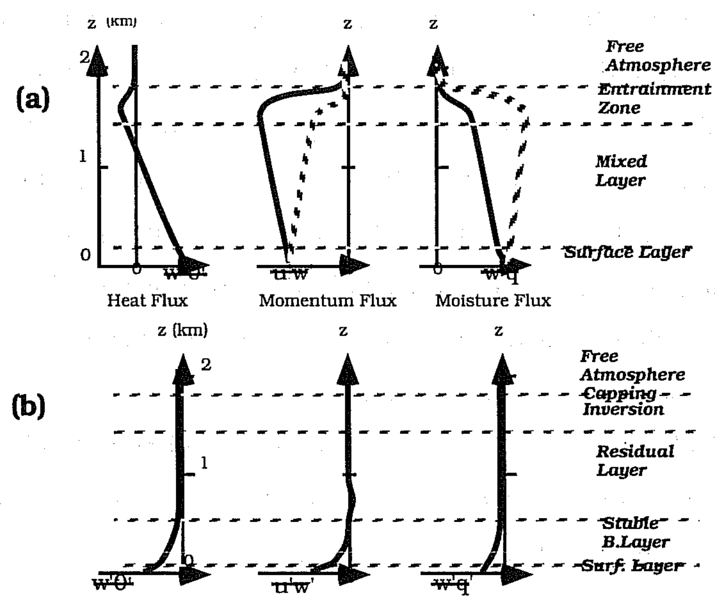

Fig. 9. Turbulent fluxes during (a) day and (b) night.

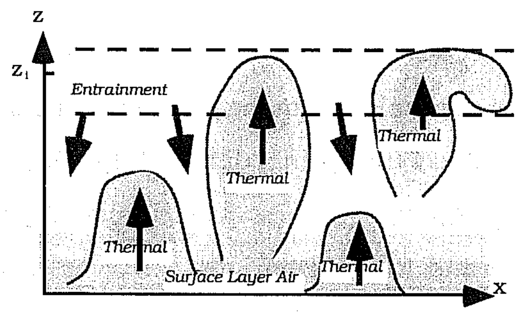

Fig. 10. Thermal structures in the convective mixed layer

Thermals with sufficient moisture can reach condensation before hitting the inversion, thereby forming a cumulus cloud (Fig 11, Schrieber et al, 1994). The latent heat released during condensation can give clouds sufficient buoyancy in some situations to puncture the inversion at the top of the ABL, to continue to grow into the FA and vent tracers from the ABL.

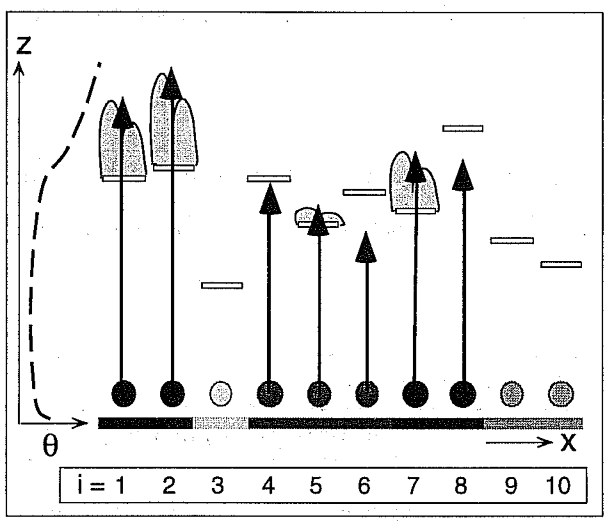

Fig 11. Air parcels (circles) rise to heights depending on their buoyancy (see dashed sounding), and form clouds depending on their individual lifting condensation levels (horizontal bars).

3.2. **Stable boundary layer**

Turbulence can be weak, patchy, and sporadic an night, making the SBL extremely difficult to parameterize. The time scale for surface information of tracers to reach the top of the SBL from the surface can be longer than the duration of the night at mid latitudes. Other factors such as gravity (buoyancy) waves, drainage winds, and direct radiative cooling of the air, and subsidence are important and difficult to parameterize. Static stability tends to suppress vertical turbulence and mixing, causing strong anisotrophy.

3.3. Residual layer

While this layer is temporarily decoupled from the ground by the intervening SBL, the wind in it can change direction and can advect tracers in a different direction and speed than the tracers in the SBL. Thus, the following day when a new mixed layer forms, this ML includes SBL air originating from one location and RL air from others. Over several days, this amounts to a splitting and merging of trajectories below the main inversion at the top of the mixed layer. Turbulence in the RL is generally weak and isotropic.

3.4. Neutral boundary layer

For windy overcast conditions, turbulence is generated continuously day and night by wind shear. This creates the neutral boundary layer. The depth of the neutral boundary layer is still controlled by the temperature inversion discussed in section 1. Turbulence is slightly anisotropic with a bit more energy in the horizontal along-wind direction than the other directions.

4. SCALING VARIABLES

Constraints such as ABL depth and imposed forcings such as surface fluxes control the turbulent nature of the boundary layer. These controls are described by scaling variables (Table 1).

Table 1. Scaling variables for the boundary layer.

Name	Symbol and Formula
Length	
height above ground	z
mixed-layer depth	z_i
Obukhov length	$L = \dfrac{-T_v \cdot u_*^3}{k \cdot g \cdot \left[\overline{w'\theta_v'}\right]_{sfc}}$
Velocity	
friction velocity	$u_* = \left[\overline{u'w'}_{sfc}^2 + \overline{v'w'}_{sfc}^2\right]^{1/4} = \sqrt{\tau_{sfc}/\rho}$
turbulence kinetic energy (TKE) scale	$e = \sqrt{TKE/m} = \left[0.5 \cdot \left(\overline{u'^2} + \overline{v'^2} + \overline{w'^2}\right)\right]^{1/2}$
Deardorff convective velocity	$w_* = \left[\frac{g \cdot z_i}{T} \overline{w'\theta'}_{sfc}\right]^{1/3}$
Buoyancy velocity	$w_B = \left[\frac{g \cdot z_i}{T}(\theta_{v\ skin} - \theta_{v\ ML})\right]^{1/2}$
Temperature	
surface-layer temperature scale	$\theta_{*\ SL} = -\overline{w'\theta_v'}_{sfc} / u_*$
mixed-layer temperature scale	$\theta_{*\ ML} = \overline{w'\theta_v'}_{sfc} / w_*$

5. HOT TOPICS

Boundary-layer meteorology is still in its infancy. While a fairly complete picture is now available for the ABL in fair weather over a horizontally-homogeneous surface, our knowledge is incomplete for some of the more useful scenarios. Some of the topics of current research interest are:

- Mesoscale modulation of the ABL
- ABL over heterogeneous landscapes
- Aggregation within a GCM grid cell
- ABL over irregular topography
- Frontal & cyclone boundary layers
- Cloud-topped boundary layers

6. REFERENCES

Ball, F.K., 1960: Control of inversion height by surface heating. *Quart. J. Roy. Meteor. Soc.*, **86**, 483-494.

Deardorff, J.W., 1972: Numerical investigation of neutral and unstable planetary boundary layers. *J. Atmos. Sci.*, **29**, 91-115.

Deardorff, J.W., 1974: Three-dimensional numerical study of turbulence in an entraining mixed layer. *Bound.-Layer Meteor.*, **7**, 199-226.

Deardorff, J.W. and G.E. Willis, 1985: Further results from a laboratory model of the convective planetary boundary layer. *Bound.-Layer Meteor*, **32**, 205,236.

Hibbard, M.F. and B.L. Sawford, 1994: A saline laboratory model of the planetary convective boundary layer. Bound.-Layer Meteor., 67, 229-250.

Schrieber, K, R. Stull and Q. Zhang, 1994: Boundary-layer cumulus: Part 1 - distributions of buoyancy vs LCL over a heterogeneous surface. (submitted to J. Atmos. Sci.)

Stull, R.B., 1988: *An Introduction to Boundary Layer Meteorology*. Kluwer Academic Publ., Dordrecht, The Netherlands. 666 pp.

Stull, R.B., 1991: Static stability — an update. *Bull. Amer. Meteor. Soc.*, **72**, 1521-1529.

Stull, R..B., 1994: *Meteorology Today for Scientists and Engineers*. West Educational Publ., Minneapolis. 350 pp.

Williams, A.G. and J.M. Hacker, 1992: The composite shape and structure of coherent eddies in the convective boundary layer. Bound.-Layer Meteor., 61, 213-246.