Mesoscale Meteorology: The Planetary Boundary Layer

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Introduction

The *planetary boundary layer*, sometimes referred to as the *atmospheric boundary layer*, is a layer of finite depth over which the Earth's surface is directly communicated to the atmosphere above. This is manifest as friction, slowing the horizontal wind; a rigid lower boundary, such that vertical motion is zero right at the surface; and surface heat and moisture exchange. The surface's direct effects upon the atmosphere are found only within the lowest few millimeters to centimeter above ground within what is known as the *viscous sublayer*. Heat and momentum exchange with the surface – as dry convection and friction, respectively – within this layer excites *turbulence*, the vertical depth of which defines the planetary boundary layer's vertical extent. Turbulence excited by dry convection is known as *buoyancy-driven turbulence*, whereas that driven by friction and its associated vertical wind shear is known as *mechanically-driven turbulence*. The *free atmosphere* resides atop the planetary boundary layer, so named because it is free from the surface below.

Covariances, Fluxes, and Flux Divergences

In statistical terms, the variance of some variable *x* is defined as the expected value – i.e., mean – of the squared deviation of *x* from its mean $\overline{x}$:

$$\sigma^2 = \overline{\left(x - \overline{x}\right)^2} = \overline{\left(x - \overline{x}\right)\left(x - \overline{x}\right)}$$

Likewise, the *covariance* between two variables *x* and *y* is defined as the mean of the product of the deviations of *x* and *y* from their respective means:

$$\mathrm{cov}(x, y) = \overline{\left(x - \overline{x}\right)\left(y - \overline{y}\right)} = \overline{x' y'}$$

Here, $x' = x - \overline{x}$ and $y' = y - \overline{y}$. In meteorological parlance, terms such as the above are known as *fluxes*. This definition will become a bit clearer shortly.

While the primitive equations apply to all scales of motion, most processes within the planetary boundary layer occur on spatiotemporal scales below those which we routinely observe (or resolve within numerical model analyses and forecasts). Instead, we apply a technique known as *Reynolds averaging* to the primitive equations such that they are formally applicable only on the observable or resolvable scales.

In Reynolds averaging, each variable is expressed in terms of mean and perturbation components. The mean component is taken to be that which can be observed or resolved, wherein the mean is usually considered to be a time and/or space average. The perturbation component is taken to be that which is unobserved or unresolved, though this should not be conflated with unimportant!

Generally, Reynolds averaging proceeds by substituting for each variable, Reynolds averaging each primitive equation, and applying Reynolds postulates to simplify. The resulting equations, given by (4.19)-(4.23) in the course text, are identical to their original forms with two exceptions.

First, all forcing terms have overbars, indicating them as time and/or space averages. For example, in numerical weather prediction models, the time average is between model time steps (seconds to minutes), while the space average is over the model grid box (usually kilometers on a side).

Second, there are several added covariance terms in the new formulation. These terms, involving partial derivatives in *x*, *y*, and *z* (or *p*, depending on the vertical coordinate), are known as *flux divergence* terms. Why are these known as flux divergence terms? Let us consider the vertical heat flux divergence term that appears in the Reynolds-averaged thermodynamic equation:

$$-\frac{\partial}{\partial z}\left(\overline{w'\theta'}\right)$$

Using a centered finite difference approximation, the vertical heat flux divergence is given by:

$$-\frac{\partial}{\partial z}\left(\overline{w'\theta'}\right)=\frac{\overline{\left(w'\theta'\right)_{z+1}-\left(w'\theta'\right)_{z-1}}}{2\Delta z}$$

Consider the case where *z+1* is at the top of some box, while *z-1* is at the bottom of some box, so that *2*Δ*z* is simply the box's height. This expression indicates that the vertical heat flux divergence is equal to how *w'θ'* changes between the top and bottom of the box. If the two are equal, then this term is equal to zero. It is only non-zero if the vertical heat flux is different at the box's top than at the box's bottom; i.e., vertical heat flux is converging into or diverging out of the box.

While the Reynolds-averaged primitive equations contain both horizontal (*x*, *y*) and vertical (*z*) flux divergence terms, in general vertical fluxes and their divergences associated with overturning eddies are an order of magnitude larger than the horizontal fluxes and their divergences associated with horizontal eddies. Thus, much of our discussion focuses on vertical flux divergence forcings. Primitive equation turbulent flux terms quantify the *mean* effects of turbulent heat, moisture, and momentum transport by eddies within the planetary boundary layer.

Surface Energy Balance

At the surface, there exists a balance between net radiation, sensible heating between the surface and atmosphere (driven by dry convection), latent heating between the surface and atmosphere, and heat transport from the surface to the sub-surface, i.e.,

$$R_n = Q_h + Q_e + Q_g$$

Here, R_n is net radiation; Q_h is the sensible heat flux; Q_e is the latent heat flux, and Q_g is the ground heat flux. The ground heat flux is typically an order of magnitude smaller than sensible and latent heat fluxes. Net radiation is given by the balance between incoming shortwave radiation from the sun, outgoing longwave radiation from the surface, and incoming longwave radiation emitted by clouds and the atmosphere. This net radiation is influenced by surface and atmosphere conditions.

The sensible and latent heat fluxes can be expressed mathematically as:

$$Q_h = -\rho c_p c_h \left\|\overline{\mathbf{v}_{10m}}\right\|\left(\overline{T_{2m}} - \overline{T_{sfc}}\right)$$

$$Q_e = -\rho l_v c_e \left\| \overline{\mathbf{v}_{10m}} \right\| \left(\overline{r_{v-2m}} - \overline{r_{v-sfc}} \right)$$

Here, ρ is density, c_p is the specific heat at constant pressure, l_v is the latent heat of vaporization, c_h and c_e are exchange coefficients for heat and moisture respectively, $\left\| \overline{\mathbf{v}_{10m}} \right\|$ is the mean 10-m wind speed, T is temperature, and r_v is mixing ratio. Typically, c_h is assumed to be equal to c_e, and each have values on the order of 0.001 to 0.01. Their values increase as static stability decreases, roughness increases, or wind measurement height decreases.

For both sensible and latent heat flux, the air loses heat and/or moisture to the surface if it is warmer and/or moister than the surface. The air gains heat and/or moisture from the surface if it is cooler and/or drier than the surface. In other words, sensible and latent heat flux magnitudes depend upon surface (land or water) temperature and moisture content, respectively, and both depend upon wind speed. These quantities are each highly variable in space and time.

Sensible heat flux is related to atmospheric heating from below. The atmosphere is transparent, or nearly so, to incoming shortwave radiation from the sun. Instead, daytime heating of the planetary boundary layer is accomplished by sensible heating from the underlying surface (which absorbs a fraction of the incoming shortwave radiation, the exact value of which is the albedo) and its vertical transport. It has large positive values during the day and warm season as well as over surfaces with high heat conductivity (e.g., sand). It is negative at night.

Latent heat flux is related to phase changes of water: e.g., evaporation of soil moisture or surface water, transpiration by vegetation, or melting and sublimation of frozen surfaces. Where there is little surface water, the latent heat flux is near-zero (or even negative). It has large positive values over warm bodies of water (such that r_{v-sfc} is large) or hot, wet soils. It is generally negative over land during the local nighttime hours.

The *Bowen ratio* represents the ratio between the sensible and latent heat fluxes. The Bowen ratio is smallest ($O(0.1)$) over oceans and wet land surfaces such as marshes and jungles. It is largest (>1) in deserts and drought-ridden locations. The Bowen ratio is related to the strength of vertical mixing within the boundary layer: larger Bowen ratios are associated with stronger, deeper vertical mixing. In other words, more of the net radiation at the surface goes into sensible heating, so that dry convection and thus thermally-driven turbulent eddies are comparatively intense.

Boundary Layer Structure and Evolution

The planetary boundary layer has pronounced structural differences between day and night. During the day, sensible heating forces dry convection and buoyancy-driven turbulence. Mechanically-driven turbulence associated with vertical wind shear, whether resulting from friction or otherwise, is also often present. This turbulence drives overturning eddies, which can be as shallow as a few centimeters or as deep as the entire planetary boundary layer. Eddies act to *mix*, or *homogenize*, material properties (potential temperature, mixing ratio, and momentum) in the planetary boundary layer. Consequently, the planetary boundary layer is often characterized as a *mixed layer*.

Following sunrise, the ground warms due to the absorption of incoming solar radiation. This results in a sensible heat flux to the atmosphere, warming the layer in contact with the surface (the *surface layer*). A superadiabatic lapse rate results, which itself triggers turbulent vertical eddy formation. These eddies mix over their depth. The upward branch of these eddies becomes negatively buoyant upon reaching the top of the boundary layer; there, *entrainment* from the free atmosphere occurs within a narrow *entrainment zone*, which helps to gradually deepen the boundary layer with time. This process continues through the day, slowing after peak heating and terminating near sunset.

After sunset, buoyancy-driven turbulence ceases, and the increasing stability of the near-surface layer resulting from longwave radiation-driven cooling can result in it *decoupling* from the free atmosphere. The increasing near-surface stability helps mitigate mechanically-driven turbulence, with the termination of nearly all turbulence leading to the development of a shallow nocturnal boundary layer and surface-based radiation inversion. However, the large-scale horizontal pressure gradient often ensures that winds do not completely still – and thus mechanically-driven turbulence does not entirely cease – at night. In fact, in cases where the meso- to synoptic-scale vertical wind shear is relatively large, turbulence may not cease and the characteristic nocturnal boundary layer may not develop. Further, cloud cover may trap part of the outgoing longwave radiation, mitigating nocturnal boundary layer development on cloudy relatively to clear nights. It stands to follow that nocturnal boundary layers are most stable on clear, calm, long nights, particularly when the ground is an efficient radiator of longwave radiation (e.g., when snow-covered). Above the nocturnal boundary layer, the remnant mixed layer – or *residual layer* – remains. It is in this layer where the nocturnal low-level jet discussed in a previous lecture develops.

The daytime planetary boundary layer can be split in two: a *surface layer*, generally representing the lowest 10% or so of the planetary boundary layer, and a mixed layer. In the surface layer, heat, moisture, and momentum exchange with the surface are sufficiently strong so as to overcome the homogenizing effects of vertical mixing. Here, potential temperature and water vapor mixing ratio decrease with height and wind speed increases with height. Within the mixed layer, each quantity is approximately constant, or nearly so, with height. Since vertical mixing reduces the potential temperature at planetary boundary layer top from what it would otherwise be, there exists a layer of relatively high static stability atop the planetary boundary layer. This is the entrainment zone and is sometimes associated with a capping inversion.

The vertical wind speed profile in the surface layer is known as a *log wind profile*, as it takes the general form:

$$\bar{u} = \frac{u*}{k} \ln\left(\frac{z}{z_0}\right)$$

Here, $u_*^2 = \left\|\overline{u'w'}\right\|_{sfc}$, where $u*$ is the friction velocity and depends on the surface drag magnitude, z_0 is the roughness length, or the height at which $\bar{u} = 0$, and k~0.4 is von Karman's constant. Friction slows the wind in the boundary layer, reducing the Coriolis force magnitude. The resulting balance between the horizontal pressure gradient, Coriolis, and frictional forces in the boundary layer, or *antitriptic wind balance*, leads to wind blowing across isobars toward low pressure.

In the nocturnal boundary layer, potential temperature, water vapor mixing ratio, and wind speed all increase with height. Potential temperature increases with height due to the longwave radiation-driven surface cooling at night. Water vapor mixing ratio decreases with height as surface cooling cools the air temperature to the dew point, thus resulting in dew formation (i.e., water vapor is lost to condensation into dew). In the case where dew does not form, water vapor mixing ratio may not decrease as rapidly with height in the nocturnal boundary layer. Wind speed decreases with height because of the constraint that it must go to zero right at the surface. Each variable decreases most rapidly with height near the surface, then less rapidly until the top of the nocturnal boundary layer. Atop the nocturnal boundary layer resides the residual zone, where vertical profiles of potential temperature and mixing ratio are similar to the daytime. Wind speed increases with height in this layer, however, due to the inertial oscillations that result from the loss of friction within this layer.

The vertical potential temperature, water vapor mixing ratio, and wind speed profiles can be used to infer the sign and relative magnitude of the turbulent vertical flux vertical profiles. Consider the daytime case. Sensible heating ensures that overturning eddies mix relatively warm air upward – i.e., $w' > 0$ and $\theta' > 0$. In the entrainment zone, this upward-mixed air is no longer relatively warm compared to its surroundings – i.e., $w' > 0$ and $\theta' < 0$. Latent heating (assuming a comparatively moist underlying surface) ensures that overturning eddies mix relatively moist air upward – i.e., $w' > 0$ and $r_v' > 0$. In the entrainment zone, this upward-mixed air is still relatively moist compared to its surroundings. Finally, friction ensures that overturning eddies mix relatively low momentum air upward – i.e., $w' > 0$ and $u' < 0$. In the entrainment zone, this upward-mixed air is still relatively low momentum compared to its surroundings.

Now, consider the nighttime case. Due to longwave radiation-induced cooling, overturning eddies would mix relatively cool air upward – i.e., $w' > 0$ and $\theta' < 0$. The reduction in water vapor mixing ratio at the surface accompanying this cooling ensures that overturning eddies would mix relatively dry air upward – i.e., $w' > 0$ and $r_v' < 0$. Finally, friction still ensures that overturning eddies would mix relatively low momentum air upward – i.e., $w' > 0$ and $u' < 0$.

The governing equations for u, v, θ, and r_v each relate their local time tendencies to the negative of the vertical flux divergence for each quantity. Thus, vertical flux divergence decreases the value of a given quantity at the level/location where its time tendency is evaluated, while vertical flux convergence increases the value of a given quantity at the level/location where its time tendency is evaluated. In general, the daytime planetary boundary layer warms, moistens, and slows due to their respective flux divergences; the nocturnal boundary layer cools, dries, and slows due to their respective flux divergences.

Boundary Layer Convection and Horizontal Convective Rolls

In general, convection in the daytime planetary boundary layer is dry convection. However, if it is sufficiently strong, shallow moist convection may result. Such convection can range from cumulus cloud streets on small scales to cellular convection on larger scales. It can, but does not necessarily always, precipitate. There are three primary types of this shallow convection with which we are primarily concerned: open cell convection, closed cell convection, and horizontal convective rolls.

Open cell convection is characterized by walls of clouds surrounding open, cloudless areas. Closed cell convection is characterized by cloud-filled areas surrounded by rings of cloud-free conditions. In the former, descent is found within each cell; in the latter, ascent is found within each cell. Both open and closed cell convection are primarily found over oceans. Open cell convection is favored when the underlying water is relatively warm compared to the overlying air and/or in the presence of large-scale descent. Closed cell convection is favored when the underlying water is relatively cold compared to the overlying air and/or in the presence of large-scale ascent. Both have vertical scales approximately $1/10^{th}$ to $1/30^{th}$ their horizontal scales. Closed cell convection is prevalent over western mid-latitude oceans on the periphery of subtropical anticyclones, where ocean surface temperature is cool relative to the overlying air.

Horizontal convective rolls are counterrotating vortices in the vertical plane. Their depth spans the planetary boundary layer, while their width can range from 3-10x their depth. In general, the width of a horizontal convective rolls increases with planetary boundary layer depth (and thus through the day); rolls are wider when the planetary boundary layer is deeper. The ascending branch of a horizontal convective roll is a favored location for cloud development in what are known as cloud streets; here, potential temperature, water vapor mixing ratio, and planetary boundary layer depth are all comparatively high compared to the descending branch of a horizontal convective roll.

There are several different instabilities that have been proposed as leading to horizontal convective rolls: thermal, inflection point, and parallel instability. Thermal instability is a function of the local buoyancy; i.e., once it becomes sufficiently large, organized convective rolls (rather than just local turbulence) may form. As entrainment reduces buoyancy (e.g., by the infiltration of comparatively cool air), and vertical wind shear is a leading cause of entrainment, thermally-driven horizontal convective rolls generally align parallel to the local vertical wind shear vector so that entrainment is minimized along the resulting cloud streets.

Inflection point instability results from a planetary boundary layer vertical wind profile containing a wind speed maximum (i.e., a vertical wind shear maximum) within the profile. This results in horizontal convective rolls aligned normal to the vertical wind shear vector or 10-20° to the left of the geostrophic wind within the boundary layer. Parallel instability is similar, but unlike inflection point instability, it requires non-zero Coriolis and frictional forces. It generally results in horizontal convective rolls aligned 10-20° to the right of the geostrophic wind within the boundary layer. In general, however, observed horizontal convective rolls likely result primarily from thermal and inflection point instabilities and are aligned between 0-30° of the mean boundary layer wind.