Mesoscale Meteorology: Drylines and the Low Level Jet

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Drylines

Overview

The *dryline* is the name for the boundary that separates hot, dry air of continental tropical origin, from warm, moist air of maritime tropical origin. In mid-latitudes, drylines are commonly found leeward of elevated terrain; in the United States, this characterizes the western Great Plains, with the Rocky Mountains to the west and flatter terrain closer to the Gulf of Mexico to the east. Drylines are most common in the warm season, when insolation and vertical mixing are strongest. In the United States, drylines are typically oriented parallel to upwind mountain ranges on the synoptic-scale; on the meso- and microscales, however, dryline bulges are frequently observed.

Drylines are characterized primarily by sharp horizontal gradients in moisture, whether in terms of dew point temperature or water vapor mixing ratio; gradient sharpness varies from case to case, with gradient magnitudes ranging from 10 K per 1-100 km. The dryline is placed on the leading edge of the moisture gradient. Winds ahead of a dryline are typically out of the south to southeast, whereas they are typically out of the west behind a dryline. Consequently, drylines are mesoscale regions of enhanced convergence. There is sometimes, but not always, a mesoscale pressure minimum found along or immediately rearward of the dryline. It is generally clear apart from possible cirrus clouds associated with an upper tropospheric jet behind a dryline, whereas cumulus, stratocumulus, or even stratus clouds can be found ahead of a dryline.

During the local daytime hours, temperature is slightly higher behind versus ahead of the dryline; during the local nighttime hours, the opposite is typically true. Both the atmosphere and, generally, underlying soil behind a dryline are drier than their counterparts ahead of a dryline. Water has a higher specific heat capacity than air, meaning that it warms and cools less rapidly than air. Thus, drier air behind the dryline cools more rapidly than the moister air ahead of the dryline. The dry, sandy soils behind the climatological dryline aid this process; i.e., there is typically minimal latent heat flux from the soil to the air behind the dryline to counteract this cooling. Since warm, moist air is approximately as dense as hot, dry air (hot air is less dense than warm air, but dry air is denser than moist air), the lack of a cross-dryline density gradient implies that drylines are not fronts.

Vertical mixing in the daytime planetary boundary layer is stronger and deeper behind the dryline than it is ahead of the dryline. This can be understood using the *Bowen ratio*, or the ratio between the sensible and latent heat fluxes. Bowen ratio is directly correlated with vertical mixing intensity. For a given amount of insolation, arid soils to the west will heat more effectively than moister soils to the east. This results in larger sensible heat flux west of the dryline versus east. Further, the arid soils to the west will transfer less latent heat from soil to air than the moister soils to the east. Thus, larger Bowen ratio is found west of the dryline, with smaller Bowen ratio to the east.

Vertical mixing homogenizes potential temperature (θ = conserved), water vapor mixing ratio (r_v = conserved), and momentum (M = conserved). Thus, vertical mixing promotes a dryline boundary that is nearly upright over a depth of 1-1.5 km along its leading edge. Drylines slope quasi-

horizontally eastward over the maritime tropical air mass, the depth of which grows eastward as terrain height gets smaller away from the Rocky Mountains. Strong sensible heating of the post-dryline air mass results in the development of a mixed layer; westerly flow aloft can subsequently advect this mixed layer downstream, atop the maritime tropical air mass, resulting in the formation of an *elevated mixed layer*. Vertical soundings within the pre-dryline air mass are characterized by relatively cool, moist air near the surface, a capping inversion or *cap* atop the planetary boundary layer, and (in some cases) an elevated mixed layer atop the cap. The cap promotes potential instability, with moist, high θ_e air beneath dry, low θ_e air, high static stability, and CIN; nearly dry adiabatic lapse rates with elevated mixed layers promote large CAPE.

Controls on Dryline Strength

Dryline strength is directly correlated to horizontal confluence magnitude: a small angle between the axis of dilatation and isodrosotherms promotes larger dew point temperature and water vapor mixing ratio gradients across the dryline. On the synoptic-scale, confluence is large in the presence of lee cyclogenesis downwind of the Rocky Mountains. Lee cyclogenesis occurs in response to upstream shortwave troughs in the middle-to-upper troposphere; the vertical stretching of an air column as it descends on the lee side of the mountains increases its relative vorticity. Lee troughs advect hot, dry air from the southwest United States and northern Mexico northeastward; with a downstream subtropical anticyclone, they help advect warm, moist air from the Gulf of Mexico northwestward. As we might expect, the gradient between these air masses defines the dryline.

On the mesoscale, confluence can be enhanced by cross-dryline differential heating. Temperature is slightly higher behind the dryline during the local daytime hours, resulting in a weak horizontal thickness and, by extension, near-surface pressure gradient across the dryline, with relatively low pressure behind the dryline and relatively high pressure ahead of it. This enhances flow toward the dryline ahead of the dryline itself; e.g., from relatively high to relatively low pressure. Behind the dryline, deep vertical mixing homogenizes momentum; for westerly winds aloft, this can result in enhanced westerly momentum mixed down to the surface behind the dryline. The cross-dryline confluence that results can locally enhance dryline strength.

Dryline Propagation

In the absence of strong synoptic-scale forcing, dryline propagation is a function of vertical mixing within the daytime planetary boundary layer. The location of the dryline can be viewed as where the top of the maritime tropical air mass intersects the gently-sloping terrain of the Great Plains. From west to east, the depth of the maritime tropical air mass grows as the terrain height decreases. Thus, the depth to which vertical mixing must extend in order to mix out the maritime tropical air mass is shallower to the west and deeper to the east. Further, the shallower depth to the maritime tropical air mass along the dryline's leading edge ensures greater surface heating (less absorption of insolation by atmospheric water vapor) there, increasing sensible heat flux and vertical mixing depth and intensity. Consequently, drylines *mix*, or propagate, eastward during the local daytime hours. Dryline eastward propagation thus is directly correlated with the maximum vertical mixing depth in the pre-dryline air mass.

During the local nighttime hours, dry post-dryline air cools more rapidly than the maritime tropical air mass found ahead of the dryline. This creates a radiation inversion, terminating vertical mixing. With low pressure to the west and high pressure to the east, flow near the surface with an easterly component advects higher moisture air westward, resulting in a dryline that retreats westward.

Locations where the dryline locally propagates further eastward are *dryline bulges*. Dryline bulges typically form in one of two ways. The first is driven by horizontal variation in sensible heating, such as that associated with horizontal soil moisture gradients. Sensible heat flux is higher over drier soils, thus promoting locally deeper mixing and increased likelihood of locally mixing out the maritime tropical air mass. Horizontal insolation gradients, such as associated with cloud cover on one or both sides of the dryline, can also contribute to horizontal variation in sensible heating and thus dryline bulges.

The second is driven by upper tropospheric jet streaks. Consider the case of a westerly jet streak's exit region impinging upon a dryline. If the full wind is set equal to the sum of the geostrophic and ageostrophic wind, i.e., $v = v_g + v_{ag}$, then it can be shown that:

$$\frac{du}{dt} = fv_{ag}$$

where u is the zonal component of the wind, v_{ag} is the meridional ageostrophic wind, and f is the Coriolis parameter. Zonal flow accelerations thus are associated with non-zero ageostrophic flow at jet level: from south to north for westerly accelerations and from north to south for decelerations. This promotes convergence at jet level in the left entrance and right exit, and divergence at jet level in the right entrance and left exit regions, of the jet streak. From continuity, this implies ascent in the right entrance and left exit and descent in the left entrance and right exit regions.

In the left exit region, forcing for ascent promotes a local reduction in sea-level pressure and thus mesoscale cyclogenesis. The resultant lower tropospheric cyclonic flow, with westerly flow to the south, can be viewed as a means of dryline bulge formation. Forcing for descent in the right exit region can promote stronger insolation, stronger sensible heat flux, and deeper vertical mixing, the latter of which promotes locally-enhanced downward transport of westerly momentum. This, too, can be viewed as a means of dryline bulge formation.

On the microscale, the intersection of horizontal convective rolls in the pre-dryline environment with the dryline can result in locally-enhanced convergence and microscale cyclogenesis. This can result in localized dryline bulges via a similar mechanism to that described on the mesoscale above.

Deep, Moist Convection and the Dryline

Drylines are a preferred locus for warm-season thunderstorm formation along the Great Plains. There are two particular locations along drylines where convection initiation is favored: near the *triple point*, or intersection of a dryline, trailing cold front, and leading warm or stationary front, and along the intersection of horizontal convective rolls in the pre-dryline environment with the dryline itself. Each are associated with locally-enhanced lower tropospheric convergence, on the mesoscale in the former case and the microscale in the latter case, and can also locally modify the lower tropospheric vertical wind shear.

Convection along the dryline tends to initiate during the local late afternoon to early evening hours. In the local morning, the pre-dryline air mass is capped by the elevated mixed layer. In cases where convection is able to initiate along the dryline, cap strength decreases through the day primarily as a result of lift across scales: synoptic-scale forcing ahead of an upstream shortwave trough, mesoscale forcing with convergence along the dryline, and microscale forcing with the ascending branch of horizontal convective rolls intersecting the dryline or, in some cases, gravity waves that are triggered by horizontal convective roll updrafts impinging upon the cap. The evolution of the thermodynamic profile atop the cap can also contribute to enhanced or diminished potential for convection initiation along the dryline. Layer lifting that results from the above forcings gradually weakens the cap; if the cap can be sufficiently weakened (e.g., erosion of negative buoyancy or CIN), potential instability may be released (e.g., conversion of CAPE into updraft kinetic energy) and convection initiation may occur. This generally occurs locally, rather than on larger-scales.

The Low-Level Jet

Overview

There are three potential interpretations of the term *low-level jet*, generically implying a corridor of enhanced lower tropospheric flow (typically poleward, and maximized between 850-700 hPa). In reality, a low-level jet in the central United States often results from a combination of all three forcings to varying degrees. All are often found along or immediately ahead of a dryline during the warm-season.

The first arises from geostrophic balance, where it can be shown that:

$$\mathbf{v}_g = \frac{1}{f}\mathbf{k} \times \nabla \Phi$$

where Φ is the geopotential (equal to gz) and f is the Coriolis parameter. From this, geostrophic wind speed is directly proportional to the magnitude of the horizontal geopotential gradient: where the horizontal geopotential gradient is large, such as between a cyclone and anticyclone, so too is the geostrophic wind speed. A corridor of enhanced lower-tropospheric wind speed aligned with the horizontal geopotential gradient can be viewed as a low-level jet. The primary warm conveyor belt of a mid-latitude cyclone can be viewed as one such low-level jet.

The second arises from forcing associated with an upper-tropospheric jet streak. For drylines, we considered a westerly jet streak impinging upon a dryline. The ageostrophic flow associated with that jet streak can be generalized to a jet streak of any horizontal orientation: from left to right in the entrance region and right to left in the exit region. This generally results in poleward flow at jet level in a jet's exit region and equatorward flow at jet level in a jet's entrance region.

If the static stability is sufficiently low (e.g., a small change in potential temperature with height, such as with an elevated mixed layer), this ageostrophic flow can extend over greater vertical depth than if the static stability were higher. Here, a poleward-directed ageostrophic wind component in the jet exit region may develop atop the planetary boundary layer – defined by the top of the mixed layer during local daytime and top of the nocturnal inversion during local nighttime hours – with compensating ageostrophic flow of opposite direction confined to immediately near the surface.

The third arises from inertial oscillations of the flow, as augmented by sloping terrain. During the day, vertical mixing communicates the effects of friction upward through the planetary boundary layer, reducing the wind speed at the top of the planetary boundary layer from that which would be expected under geostrophic balance. Thus, the relevant force balance is between the horizontal pressure gradient, Coriolis, and frictional forces, and the wind speed is subgeostrophic. After local sunset, insolation terminates, the surface cools, and a radiation inversion develops. This increases near-surface static stability, terminating vertical mixing, and decoupling the surface layer from the middle to upper reaches of the daytime planetary boundary layer.

This disrupts the aforementioned force balance between the radiation inversion top's and remnant mixed layer's bottom, as friction becomes negligible. The atmosphere attempts to restore balance (here, geostrophic balance) through an acceleration of the ageostrophic flow (i.e., supergeostrophic flow develops), resulting in an inertial oscillation of the wind about its geostrophic balanced state and a nocturnal lower tropospheric wind maximum. Diurnal variation in heating on sloped terrain leeward of the Rocky Mountains can augment this maximum. We now wish to consider inertial oscillation development in a simplified mathematical framework.

Mathematical Development

The derivation below largely follows that from Section 4.7.3 of the course textbook with only a few minor additions. We start with the horizontal momentum equations valid on constant height surfaces. We neglect friction given our focus on when friction is unimportant. Thus, we obtain:

$$\frac{du}{dt} = -\frac{1}{\rho}\frac{\partial p}{\partial x} + fv \qquad\qquad \frac{dv}{dt} = -\frac{1}{\rho}\frac{\partial p}{\partial y} - fu$$

Let the full wind be equal to the sum of its geostrophic and ageostrophic components, i.e., $v = v_g + v_{ag}$ and $u = u_g + u_{ag}$. Substituting into the horizontal momentum equations, we obtain:

$$\frac{du_g}{dt} + \frac{du_{ag}}{dt} = -\frac{1}{\rho}\frac{\partial p}{\partial x} + f\left(v_g + v_{ag}\right)$$

$$\frac{dv_g}{dt} + \frac{dv_{ag}}{dt} = -\frac{1}{\rho}\frac{\partial p}{\partial y} - f\left(u_g + u_{ag}\right)$$

For now, we wish to assume that the background environment is otherwise unchanging. In other words, we wish to assume that the background geopotential field – and thus the geostrophic wind – is constant with time following the motion. Thus, the first left-hand-side terms in each equation above can be neglected, resulting in:

$$\frac{du_{ag}}{dt} = -\frac{1}{\rho}\frac{\partial p}{\partial x} + f\left(v_g + v_{ag}\right)$$

$$\frac{dv_{ag}}{dt} = -\frac{1}{\rho}\frac{\partial p}{\partial y} - f\left(u_g + u_{ag}\right)$$

By definition, the geostrophic wind components are given by:

$$u_g = -\frac{1}{\rho f}\frac{\partial p}{\partial y} \qquad\qquad v_g = \frac{1}{\rho f}\frac{\partial p}{\partial x}$$

This can be demonstrated from the horizontal momentum equations neglecting the total derivative terms altogether. If we substitute into the modified horizontal momentum equations, we obtain:

$$\frac{du_{ag}}{dt} = fv_{ag} \qquad\qquad \frac{dv_{ag}}{dt} = -fu_{ag}$$

Next, take the total derivative of the first equation and substitute with the second equation:

$$\frac{d^2 u_{ag}}{dt^2} = \frac{d}{dt}\left(fv_{ag}\right) = -f^2 u_{ag}$$

If we treat f as a constant, this equation represents a linear ordinary differential equation with a solution of the form:

$$u_{ag} = C_1 \cos ft + C_2 \sin ft$$

Here, C_1 and C_2 are constants, f is the Coriolis parameter, and t is time. If we rearrange the equation relating the total derivative of u_{ag} to v_{ag}, solving for v_{ag}, and substitute this solution, we obtain:

$$v_{ag} = \frac{1}{f}\frac{du_{ag}}{dt} = \frac{1}{f}\frac{d}{dt}\left(C_1 \cos ft + C_2 \sin ft\right) = \frac{1}{f}\left(-C_1 \sin ft + C_2 \cos ft\right) = -C_1 \sin ft + C_2 \cos ft$$

Let us consider $t = 0$, which we can consider to be the time when friction becomes unimportant. If we let $u_{ag} = u_{ag0}$ and $v_{ag} = v_{ag0}$ at this time, then we obtain:

$$u_{ag0} = C_1 \cos 0 + C_2 \sin 0 = C_1$$

$$v_{ag0} = -C_1 \sin 0 + C_2 \cos 0 = C_2$$

And thus,

$$u_{ag} = u_{ag0} \cos ft + v_{ag0} \sin ft$$

$$v_{ag} = -u_{ag0} \sin ft + v_{ag0} \cos ft$$

Using trigonometric identities (see, for example, the graphical illustration on Wikipedia, where $\alpha = \varphi_0$ and $\beta = ft$), it can be shown that:

$$u_{ag} = \left\|\mathbf{v}_{ag0}\right\| \cos\left(\psi_0 - ft\right)$$

$$v_{ag} = \left\|\mathbf{v}_{ag0}\right\| \sin\left(\psi_0 - ft\right)$$

Here, $\mathbf{v_{ag0}} = u_{ag0}\mathbf{i} + v_{ag0}\mathbf{j}$, $\left\|\mathbf{v}_{ag0}\right\| = \sqrt{u_{ag0}^2 + v_{ag0}^2}$, and φ_0 is the angle associated with the orientation of the geostrophic wind at the day-night transition (i.e., the onset of inertial oscillations).

The only non-constant in the above is time t. For a start time of $t = t_0$, $u_{ag}(t_0)$ and $v_{ag}(t_0)$ equal u_{ag} and v_{ag} at some later time $t = 2\pi/f$, since one full unit circle has length 2π. Thus, the time it would take for a full oscillation to complete is $2\pi/f$. Where $f = 2\Omega\sin\phi$, where ϕ is latitude and Ω is Earth's angular rotation rate, $2\pi/f = 2\pi/(2\Omega\sin\phi) = \pi/(\Omega\sin\phi)$, where π/Ω is the length of a half-rotation of the Earth (or 12 h). Thus, a full oscillation completes in 12 h/sin ϕ – or, for $\phi \sim 35°N$, about 7 h.

The equations defined above describe an oscillating ageostrophic wind with time, with clockwise rotation in the Northern Hemisphere about a constant geostrophic wind. As the atmosphere begins from a state of subgeostrophic flow, this oscillation defines a flow that becomes supergeostrophic through the night before returning to a subgeostrophic state prior to completing a full oscillation. The amplitude of this oscillation is directly related to the magnitude of the subgeostrophic flow at the start of the oscillation period (i.e., $\|\mathbf{v_{ag0}}\|$). Thus, a stronger nocturnal wind maximum can be expected when winds in the upper planetary boundary layer were most affected by friction (e.g., nominally when heating is strong, mixing is vigorous, and the planetary boundary layer is deep). Note that this oscillation results in varying wind speed and direction, with winds generally turning in a clockwise fashion (e.g., south to southwest) as night proceeds.

Naturally, oscillations do not continue forever: the onset of a new day results in the restoration of balance between the horizontal pressure gradient, Coriolis, and frictional forces within the boundary layer. Further, this framework is limited by the assumption that there is no response in the mass field to restore balance; e.g., the geostrophic wind is constant following the flow. What happens when the geostrophic wind is not constant during the overnight hours?

As noted previously, sloping terrain can influence this nocturnal oscillation. Consider the case of gently-sloping terrain with higher heights to the west, such as found across the Great Plains. During the local daytime hours, absorption of insolation by the ground causes temperature to be warmer along a constant height surface intersecting the terrain to the west than to the east above the terrain. The opposite is true at night: the loss of heat by radiation causes temperature to be cooler to the west than to the east.

The result of each is a zonal temperature gradient in the lower troposphere: warm west in the day, warm east at night. From thermal wind balance, this implies a southward-directed thermal wind in the daytime hours and a northward-directed thermal wind in the nighttime hours. In a simplified sense, with only meridionally-directed geostrophic flow (i.e., a zonal geopotential gradient), this implies an anomalous southward-directed geostrophic wind atop the boundary layer during the day and an anomalous northward-directed geostrophic wind atop the boundary layer at night. The latter adds to the northward-directed ageostrophic flow with the nocturnal wind maximum driven by inertial oscillations and represents a simplified manifestation of how non-constant geostrophic flow following the motion can influence the nocturnal lower tropospheric wind maximum.

Importance to Nocturnal Convection

The nocturnal low-level jet is an important contributor to the warm-season precipitation and severe weather climatology across the central United States. Consider the case of an upper tropospheric westerly jet streak across the central United States ahead of an upstream shortwave trough. By definition, there must be westerly vertical wind shear over the layer beneath the jet streak. From thermal wind balance, this implies a north-south layer-mean temperature gradient over the vertical wind shear layer with warm layer-mean air to the south of the jet. A poleward-directed nocturnal low-level jet impinging upon this baroclinic zone will be forced to ascend. In an environment of weak to non-existent CIN for elevated parcels, this can result in elevated convection initiation; in an environment of existing surface-based convection, the nocturnal jet acceleration can increase the mesoscale forcing for ascent, resulting in rapid upscale growth of discrete convective cells into mesoscale convective systems. The resulting organized convective systems propagate eastward with the large-scale flow, often persisting through the nighttime hours and weakening near sunrise.

The nocturnal low-level jet is also a corridor of rapid poleward transport of warm, moist air from locations nearer to the Gulf of Mexico. While this air does not always ascend over a baroclinic zone and result in elevated convection initiation, it can set the stage for surface-based convection initiation during the next diurnal maximum (e.g., in advance of an approaching upstream trough). Further, a low-level jet is, by definition, a local maximum in vertical wind shear beneath the jet, which can enhance or modify the near-surface hodograph, horizontal vorticity, and storm-relative helicity along its axis. This is particularly important during the day-night transition, where the cap is at its weakest, the near-surface layer has not yet stabilized, and yet the nocturnal low-level jet is beginning to accelerate – CIN is a minimum, CAPE is a maximum, and lower tropospheric vertical wind shear is increasing.