

Drylines and Convection: a Literature Review

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ABSTRACT

Drylines are similar to fronts, but only have a difference in moisture instead of temperature. This moisture difference, or gradient, can be quite sharp. Drylines move in jumps to the east and can retrograde in the evening. They are commonly associated with other boundaries, also, such as fronts.

Drylines can aid in convective initiation due to the large-scale and small-scale differences present at the boundary. Large scale differences and effects include localized turning of surface winds towards the boundary due to a pressure gradient force caused by differential heating, the presence of soil moisture, and pre-existing cloud streets. Small-scale phenomena such as horizontal convective rolls (HCRs) and mesocyclones create enhanced surface moisture convergence and control where updrafts can force air parcels to their LFC to trigger deep convection. However, if large-scale subsidence from a longwave or shortwave ridge, large amounts of CIN, or a maximum in upward vertical velocity located too far below the LCL or LFC are present, then convective initiation can still be rejected even if the other forcing factors are present.

1. Introduction

The main purpose of this paper is to discuss the formation, characteristics and movement of drylines. Also, this paper will discuss the role that drylines play in convection. Section 2 will provide a brief overview of the topics to be discussed. Section 3 will discuss the evolution of drylines in terms of their movement, while section 4 will discuss the formation of drylines. In section 5, the role of drylines in initiating convection will be discussed in detail, and a summary of the topics covered is included in section 6.

2. Overview

Drylines are very similar to fronts. However, instead of being characterized by a difference in temperature, they are characterized by a difference in dew point. This is the main means of defining drylines since any wind shift associated with them is usually located to the west of the dryline (Schaefer 1974). In the United States, drylines usually form on the lee side of the Rocky Mountains. They form at the convergence of warm, moist air from the Gulf of Mexico from the southeast and hot, dry air from the desert area of the southwest United States and Mexico. Also, cold air descending from the mountains warms adiabatically, and since most of the moisture has already been condensed out of it by being lifted over the mountains, it is fairly dry when it descends

from the mountains. Where these two air masses meet is defined as the dryline (Atkins et al. 1998). One other way to think of a dryline is that it is a transition zone between two different boundary layers (Karan and Knupp 2006).

Whether or not drylines aid in convective initiation is related to the large-scale and small-scale along-dryline variability that is present. Since a mesoscale circulation/dryline is frequently present, especially during the day, this variability will either add or detract from the upward motion already present. Large-scale differences include enhanced convergence at the boundary due to winds being redirected by a pressure gradient force created by differential heating (greater temperatures promote lower surface pressure behind the dryline, and cooler temperatures promote higher pressure ahead of it); soil moisture changes due to precipitation; the presence of cloud streets intersecting the dryline; cloud shields which can limit instability and increase CIN; and synoptic-scale subsidence due to the presence of a ridge. Small-scale differences include the presence of horizontal convective rolls (HCRs), which can enhance surface moisture convergence and actually deform the dryline; and mesocyclones, which can create preferred updraft locations and spawn nonsupercell tornadic circulations when collocated with an updraft core in an ongoing thunderstorm.

3. Movement and Characteristics

The movement of drylines is very different from the movement of other fronts. While the movement of other fronts is determined by the movement of the cold air, the movement of drylines is determined by the mixing of the air. The warm, dry air on the west side of the dryline overrides the moist air on the east side, thus forming a cap. Since the dry air warms faster than the moist air, a deeper boundary layer is created. The dryline advances to the east by mixing the moist air with the dry air and lowering the dew points. The dryline can actually retrograde to the west during the evening and night since it also cools off much faster due to being more dry.

One problem with drylines is being able to actually detect the passage of a dryline. One characteristic of a dryline passage is a dew point change of at least 3 degrees Celsius at a rate of at least 0.5 degrees Celsius per hour (Crawford and Bluestein 1996). Some drylines can have an associated dew point variation of up to 18 degrees Celsius. However, since convection can be associated with drylines, outflow boundaries have been known to mask their passage. Also, oscillations in dew point can be associated with waves along the dryline. In different cases of drylines, the change in dew point and the timescale over which the change occurred was very different. In some cases, the dew point changed in 10-20 minutes and in others it changed over an hour. Drylines also typically have very strong vertical wind shear that helps fire off convection. They also can have strong surface convergence associated with them. Drylines also tend to have a sharp wind shift on the western edge and generally weak flow on this side. One other interesting thing found by Karan and Knupp (2006) was that the depth of the boundary layers oscillates with the western side or dry side boundary layer being deeper than the eastern side.

The movement of drylines is not as smooth or as easily predictable as the movement of fronts. Drylines do not always move smoothly; they tend to move in jumps. The most common trend of dryline movement is for them to move smoothly

eastward during the early morning, but then to jump rapidly eastward during the afternoon. However, as they move eastward, they do not always move in a straight line but tend to form bulges in the boundary where the dryline sticks out relative to other stretches of the boundary (Hane et al. 2001). With these bulges, different sections can move faster than others and some sections can actually retrograde while other parts continue to move eastward. During the late afternoon and evening, drylines will retrograde to the west in a smooth continuous movement. According to Weiss et al. (2006), it has been found that terrain slope, surface heat fluxes, and soil moisture gradients have all been shown to have an influence on the development and movement of drylines. One interesting thing from Weiss et al. (2006) is that evidence of other boundaries forming near the dryline that act similar to the dryline was found, and in some cases those boundaries had a moisture gradient similar to that of the dryline. This could possibly explain the variability along the dryline.

As in Wakimoto et al. (2006), Doppler velocity can be used to study some of the characteristics of a dryline. Doppler velocity data was used to find evidence of low-level convergence and upper-level divergence which is consistent with the traditional characteristics of drylines. Also discovered was that the hot, dry air on the west side of the dryline flowed up and over the moist air, which also fits with the traditional image of a dryline. However, it was also discovered that there was descent 3-4 km to the east of the dryline that seems to be a part of the dryline circulation driven by the gradient of virtual potential temperature across the dryline. This secondary circulation was often characterized by an easterly region of ascent near the strong moisture gradient and a westerly component above the boundary layer on the east side of the dryline (Weiss et al. 2006). One other interesting feature from Wakimoto et al. (2006) was that there was a westward tilt with height of cyclonic vertical vorticity.

4. Formation

Weiss et al. (2006) also identified a general synoptic pattern that usually leads to the formation of drylines. These features include an eastward moving, negatively tilted long wave trough over the western U.S. at 500 mb, a weak jet across central and southern Colorado also at 500 mb, and weak 500 mb winds to the southeast of Colorado. At 850 mb, there should be south to southwest winds over Texas and Oklahoma with good moisture transport from the western Gulf of Mexico. One other thing that helps is the appearance of a strong inversion at 1200 UTC from around 850 mb to 750 mb.

5. Convective Initiation

It is generally well understood how boundaries such as cold and warm fronts aid in the initiation of convection. Drylines, however, despite being boundaries between two air masses just like warm and cold fronts, are not so simple. Although similar to cold fronts, the mesoscale circulation/updraft created by the dryline can indeed help initiate convection. Drylines, in general, possess a mesoscale circulation within 10 km of themselves that provides forcing and upward motion which promotes forced cumulus development. This forces air parcels reach their level of free convection (LFC) as well as lifting condensation level (LCL), and triggers deep convection. The mesoscale

circulation involves a prominent vertical moisture bulge in which large values of vapor mixing ratio and theta-e are forced higher up above the dryline relative to areas farther from it. This moisture comes from the surface on the moist side of the dryline. Mesoscale horizontal moisture convergence at the boundary most likely explains the upward motion (Ziegler et al. 1995; Hane et al. 1996; Ziegler et al. 1996; Ziegler and Rasmussen 1998; Grasso 1999; Richter and Bosart 2001; Xue and Martin 2006; Cai et al. 2006; Murphey et al. 2006; Holt et al. 2006). In fact, updrafts of up to 5 ms^{-1} have been modeled to be common along dryline convergence bands (Ziegler et al. 1996). How well this upward motion does in forcing air parcels upward to reach their LCL and LFC depends also on the larger scale and finer scale differences along the dryline, the effects of which will either add to the convergence and upward motion, or detract from it.

a. Large-scale along-dryline variability

Whether or not a dryline serves as a focal point for convection can depend on a great number of factors which can change not only in a distance normal to the dryline, but can also change along the dryline as well. Let's take a look at some specifics about air parcels first. Then, we will cover the regional differences and follow with a brief discussion of rejection of convection.

Air parcels need to be forced to their LCL and LFC in most cases of dryline interaction. Therefore, convection is likely as long as parcels remain in the mesoscale updraft created by the dryline circulation long enough to reach their LFC (Ziegler and Rasmussen 1998). Two sample parameters can determine the likelihood of this happening. These parameters are ratios that can be derived from radiosonde data, either by direct inspection of a skew-t, or by using values and known formulas to derive the needed quantities. One parameter is $H_{lcl} = h_{wmax}/h_{lcl}$, where h_{wmax} is defined as the height at which the maximum upward velocity in a sounding occurs, and h_{lcl} is defined as the height of the lifting condensation level. Both are measured in meters so that H_{lcl} is a dimensionless ratio. The significance of H_{lcl} is that values exceeding unity indicate an increased likelihood of air parcels to reach their LCL. The other parameter is H_{lfc} , which is defined in a similar way as H_{lcl} with the height of the level of free convection, h_{lfc} , replacing the term h_{lcl} in the formula. Again values of H_{lfc} that are greater than unity indicate an increased likelihood of air parcels reaching their LFC. Having both parameters larger than unity indicate the best chance of deep convection occurring.

Even if an air parcel is lifted to its LCL, there is no guarantee it will continue to rise, reach its LFC, and initiate deep moist convection. Many things can happen to an air parcel in the region of the LCL and LFC to prevent continued ascent, such as drying of the parcel due to entrainment from above the boundary layer (which is common with drylines since the mixing from above is one primary characteristic of their movement), and movement away from the region of the upward forcing which lifted the parcel in the first place. Even with sufficient lifting and maintenance of saturation of a lifted parcel, if sufficient amounts of CIN are present to resist upward motion, then development of convection will be temporally limited, severely restricted, or perhaps impossible (Ziegler et al. 1996). Analogous to determining the maximum updraft speed if all of the convective available potential energy (CAPE) were transformed into kinetic energy of the parcel by the formula $w_{max} = (2*CAPE)^{1/2}$, the same idea can be used with the negative

buoyant potential energy (CIN) to determine the minimum upward vertical speed a parcel would need to have after leaving the mesoscale updraft of the dryline in order to break through the layer of CIN and reach the LFC. Assuming no entrainment of cool, dry air into the parcel and that the layer of CIN begins immediately after the updraft terminates, then the velocity needed is given by $w_{\min} = (2 * \text{CIN})^{1/2}$. Obviously, the assumptions will frequently be false and this method will not apply, but this formula gives a baseline value or approximation to the strength of the forcing needed for an air parcel to break through any stable layers.

Regional effects can increase or decrease the favorability of convection by various means. It has been documented that various regional differences in the vicinity of a dryline have either suppressed or aided convection (Hane et al. 2001). One such difference is a region of moist soil juxtaposed with a region of dry soil, with the moist soil being caused by precipitation from a recent time period. Evaporation of the moisture in the soil can cause lower temperatures, which subsequently reduces surface-based instability. Another such difference is related to the fact that the drier air behind the dryline changes temperature more readily than the more moist air ahead of the dryline. During the day the temperatures behind the dryline become higher than those ahead of it. This difference in temperature causes the surface pressure behind the dryline to decrease more than that ahead of the dryline, creating a pressure gradient force which causes winds on the moist side to turn towards the boundary, towards the warmer, drier air behind the dryline. This turning towards the boundary results in enhanced surface convergence at the dryline, which can enhance lift and aid convective initiation (Hane et al. 2001; Murphey et al. 2006). The intersection of a preexisting cloud street intersecting the dryline has also been documented to aid in triggering deep moist convection (Hane et al. 1996; Hane et al. 2001). If such factors as the spread and intensity of precipitation from a previous time period, fine-scale temperature gradients, and pre-conditioning of the atmosphere through means such as cloud streets (which could possibly be related to land-use differences causing abrupt changes in surface sensible heat flux) in the region of the dryline (Hane et al. 1996) are known or can be determined, then indicating the probability and location of convective initiation near a dryline can be possible, given detailed analysis. Otherwise, we can see that it can be very difficult to predict the mesoscale influences that will determine whether or not convection initiates at a given point along the dryline.

The 3 May 1999 tornado outbreak that spawned the F5 that struck the Oklahoma City, Oklahoma area serves as another example for large-scale along-dryline variability. The supercell storm that produced that powerful tornado originated along a dryline, but it was one of only two storms (supercells A and B from Thompson and Edwards 2000) to initiate along the dryline around the time it did. The culprit for the lack of concentrated development at that time that day was a shield of high clouds, which inhibited convection. The storms eventually initiated because a small gap in the cloud shield formed which allowed sufficient surface heating to destabilize the lower troposphere enough so that the mesoscale dryline circulation could initiate the convection. The models poorly forecast this, and SPC forecasters, despite knowing an outbreak was possible, only predicted it to a small extent one day out and not as a major tornado outbreak (Thompson and Edwards 2000).

It is possible that even when the right conditions are met at the surface and air parcels are being forced up to their LCL and beyond in the region of a dryline by methods discussed previously (and also by some methods to be discussed in section 5b), convective initiation will not necessarily occur. In the presence of synoptic-scale subsidence, enough downward motion throughout a thick layer of the lower-middle troposphere can be enough to reject convective initiation. This can occur by means of adiabatic warming as the layer subsides. When this happens, a capping inversion forms, thus raising the LFC and increasing CIN. Favorable synoptic setups for this situation include the presence both long wave and short wave ridges over the dryline (Richter and Bosart 2002; Cai et al. 2006).

b. Small-scale along-dryline variability

Along with large-scale regional differences, smaller-scale elements, such as horizontal convective rolls (HCRs) and mesocyclones can help determine if, when, and where convection will initiate off the dryline.

The details of origin and structure of horizontal convective rolls are not completely understood, but they appear to be caused by some combination of convective instability caused by surface heating and from some amount of vertical wind shear. HCRs are horizontal, wavelike, rotating tubes that appear in clear air. They are confined to the boundary layer, and are counter-rotating – that is, each roll rotates in the opposite direction as its neighbors. One characteristic of HCRs is their aspect ratio which is the ratio of the width to depth of the rolls (also given as the ratio of the distance between updrafts generated by the rolls to the depth of the boundary layer) (Xue and Martin 2006a). For reference, some typical values for the aspect ratio range between 3:1 and 5:1. The axes of HCRs are oriented parallel to the boundary layer mean shear vector, or closely aligned to the boundary layer mean flow. Evidence for the existence of HCRs can appear on radar as thin lines or blotchy or patchy reflectivity from insects that get caught in the circulations caused by them (Hane et al. 2001; Markowski et al. 2006; Murphey et al. 2006). Cloud streets – lines of cumulus clouds – are also indicative of the existence of HCRs since the sections of the rolls that promote upward motion can result in updrafts of 1 ms^{-1} or greater (Xue and Martin 2006a). HCRs are also evidenced by the presence of maxima in a near-surface moisture convergence field (Ziegler et al. 1996; Xue and Martin 2006a). Those maxima are due to HCRs organizing themselves into roll convergence bands (RCBs), or narrow strips of high values of moisture convergence near the surface, interspaced with open convective cells that appear as local minima in the moisture convergence field (i.e., areas of local horizontal moisture divergence) located between successive RCBs, or essentially the regions of HCRs that contain descending air. HCRs occur in any areas as long as the mechanisms for creating them are engaged; they can occur far from any fronts or boundaries. It is when they interact with the already-present mesoscale circulation of the dryline that they become significant in determining where and when convection will fire. Where RCBs intersect the primary dryline convergence band (PDCB) (Xue and Martin 2006b), or the area toward the eastern side of the most intense moisture gradient where mostly southeasterly flow counters mostly westerly flow, is where the greatest enhancement of low-level convergence, and also upward motion, is found. Convection can form along the dryline just slightly downwind

of this intersection (Xue and Martin 2006b). The downwind motion is due to advection of the pre-precipitating cloud by mid-level winds. The pre-precipitating clouds themselves form directly over the intersection. The lifetime and intensity of the convection is related to how long air parcels remain in the region of enhanced convergence and upward motion (refer to the discussion in part (a) of this section) and related to the larger-scale dynamics encompassing that region (Cai et al. 2006). Parcels that spend more time in this area have more time to ascend to their LFC and form convection than those that spend less time there. Over time, the evolution of HCRs into RCBs and their subsequent interaction with the PDCB can cause the convergence boundary to become less linear and more wavelike in appearance (Xue and Martin 2006b). The wavelength of the deformed PDCB is related to the spacing of the RCBs. This wavelike appearance can indicate the presence of, or help create, an additional feature that can aid in convective initiation: misocyclones.

The surface flow local to RCBs near the PDCB can cause very small vertical vorticity tubes (having diameters less than 4 km), called misocyclones, to form on the PDCB between each intersection of a RCB and the PDCB (Xue and Martin 2006b). In a sort of positive feedback loop, the circulation of misocyclones can create bulges of dry, warm air to their south, and cool, moist air to their north, which can then increase the tendency for the PDCB to take on the wavelike shape, and also provide a means of very localized enhanced convergence or divergence, but the latter has not been documented. Otherwise, misocyclones do not cause any enhanced vertical motion themselves, but can have a large impact on the location of updrafts which can force air parcels to reach their LFC (Murphey et al. 2006). This is because misocyclones contain a downward-directed pressure gradient force at their core, thus suppressing any vertical motion within, but not suppressing it outside the vortex (see Fig. 16 of Murphey et al. 2006). Thus, misocyclones create preferred locations for updrafts, which will control parcel ascent. As long as a misocyclone and updraft do not become collocated before initiation, a parcel in such an updraft will be allowed to reach its LFC. In a pre-storm environment, misocyclones are usually located to the south of updraft cores with little or no superposition between them. However, after convective initiation, the updraft cores generally expand horizontally, and the misocyclones, not moving relative to the pre-storm configuration, move within the updraft cores. This is where they can really affect things. The superposition of upward vertical motion (i.e., updraft cores) and misocyclones can spawn nonsupercell tornadoes. Thus, once convection fires and the updrafts expand to cover the misocyclones, nonsupercell tornadic circulations can result. This has been documented in Murphey et al. (2006).

6. Summary

Drylines are characterized by sharp moisture gradients and tend to move in jumps, but they can also retrograde. They also tend to be associated with other boundaries such as fronts. The dryline and its interaction with these boundaries tend to lead to convection.

How well the dryline does in initiating convection depends on the large-scale and small-scale differences that enhance surface convergence and promote upward motion, with the end goal being to force air parcels to their LFC in an unstable environment.

Pressure gradient forces due to differential heating on opposing sides of the dryline, along with soil moisture and pre-existing cloud streets consist of some large-scale effects, while horizontal convective rolls and mesocyclones are two small-scale effects that enhance surface convergence and control where updrafts can exist and provide the necessary forcing of air parcels in order to initiate convection.

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