Synoptic Meteorology I: Fronts and Frontal Analysis

25 November, 2 December 2014

Introduction

In the atmosphere, we typically observe intense temperature gradients and strong winds concentrated into localized areas. The former are associated with fronts, while the latter are associated with jets. Why are fronts and jets typically linked and located in close proximity to one another? The answers, unfortunately, are not immediately obvious.

Thermal wind balance, relating the magnitude and direction of the layer-mean horizontal temperature gradient (which is often large in the vicinity of a front) to a measure of the vertical wind shear (which is often strong in the presence of a jet), helps us understand why fronts and jets are often found in close proximity to one another. However, it does not explain how either fronts or jets form or evolve, nor does it explain their fundamental structures.

Thus, we desire to address the structure, formation, and evolution of fronts and jets, phenomena across which the horizontal scale is small and, for jets, for which parcel accelerations are important. Before we do so, however, it is worthwhile to first answer two basic questions:

1) *What is a front?*

In the broadest sense, a front is a boundary between two air masses. It represents an elongated ***zone*** (*not* a finite line or local discontinuity) of locally strong horizontal temperature gradients. What do we mean by *elongated* and *locally strong*, however?

- **Elongated**: The along-front distance, on the order of 1000 km (e.g., on the synoptic scale), is much greater than the across-front distance, which is on the order of 100 km (e.g., on the mesoscale).
- **Locally strong**: The across-front horizontal temperature gradient is on the order of 10 K per 100 km, while the across-front horizontal mixing ratio gradient is on the order of 10 g kg^{-1} per 100 km. These are one order of magnitude larger than their typical synoptic-scale counterparts (e.g., the same change over a distance of 1000 km).

A frontal inversion, characterized by an increase in temperature with height (thus representing a stable situation), is found at the top of a frontal zone.

2) *What is a jet?*

A jet is an intense, narrow, quasi-horizontal current of wind that is associated with strong vertical wind shear. What do we mean by *intense*, *narrow*, and *strong*, however?

- **Intense**: The wind speed within the jet is greater than or equal to 30 m s^{-1} (~60 kt) for upper tropospheric jets and greater than or equal to 15 m s^{-1} (~30 kt) for lower tropospheric jets.

- **Narrow**: The along-jet distance, on the order of 1000 km (e.g., on the synoptic scale), is much greater than the across-jet distance, which is on the order of 100 km to 250 km (e.g., on the mesoscale).

- **Strong**: The vertical wind shear is on the order of 5-10 m s^{-1} per kilometer, or approximately five to ten times larger than its typical synoptic-scale value.

A jet may be found at any level within the troposphere. A local maximum of wind speed embedded within a jet is known as a jet streak.

Why are fronts and jets important?

- There exist large variations in meteorological conditions across both fronts and jets.

- They are typically associated with "interesting" weather, including precipitation, thunderstorms, strong winds, intense cyclone development, and so on.

- Strong vertical wind shear found with a jet is often associated with turbulence, which is a major aviation hazard.

- Fronts and jets are responsible for both horizontal and vertical transport of particulates, such as ozone and/or pollutants.

In this lecture, we wish to emphasize the observational characteristics of fronts and mature mid-latitude, synoptic-scale cyclones. In the next lecture, we will do the same for jets and jet streaks. Next semester, we will look at the dynamics of both fronts and jets, wherein the distinction between the along-front/along-jet and across-front/across-jet length scales noted above becomes crucially important.

Air Masses and their Properties

Given that a front separates two air masses, it is helpful to first briefly review the defining characteristics of air masses. An air mass is defined by its temperature and its moisture content. There exist five primary air masses: continental Arctic (cA), continental Polar (cP), continental Tropical (cT), maritime Polar (mP), and maritime Tropical (mT). The first word of each air mass designates its moisture content – continental implying dry and maritime implying moist – whereas the second word designates its temperature – Arctic implying bitterly cold, Polar implying cold or cool, and Tropical implying warm or hot.

Air masses typically form over homogeneous land surfaces where air can remain undisturbed (e.g., in place) for a relatively lengthy period of time. Fronts represent boundaries between two air masses, such as a warm and moist (mT) air mass and a cooler and drier (cP or cA) air mass. Air masses are modified primarily by one of two physical processes. Via air-sea and air-land interaction, an air mass can exchange sensible (temperature) and/or latent (moisture) heat with its underlying surface. Alternatively, as an air mass is displaced meridionally, the length of insolation that it experiences changes, thereby modifying the air mass' thermal properties.

Surface Fronts

Introduction

Fronts can be found anywhere within the troposphere. The strongest fronts extend from the surface upward to the tropopause. However, many fronts are shallower in nature and are located in either the lower troposphere or the middle to upper troposphere. A front that is strongest near the ground is known as a surface front. Such fronts generally decay in intensity with increasing altitude. Surface fronts are generally located downstream (ahead) of upper-tropospheric troughs and upstream of (behind) upper-tropospheric ridges in a region of synoptic-scale forcing for rising motion. Surface fronts are characterized by one or more of the following properties:

- A zone of strong, across-front temperature, moisture, vertical motion, and relative vorticity gradients.
- A relative minimum, compared to locations on either side of the frontal zone, of pressure.
- A relative maximum, compared to locations on either side of the frontal zone, of relative vorticity along the front.
- A zone of confluent flow along the front.
- Strong vertical wind shear along and horizontal wind shear across and along the front.
- Rapid changes in cloud and precipitation properties across the front.

Note that not all of the aforementioned characteristics of a surface front are necessarily present with any given surface boundary, nor are they all necessarily precisely co-located in space. Concordantly, these characteristics may not necessarily all move at the same rate of speed, nor may they necessarily all evolve in an identical fashion through time. We will examine many of these properties using multiple real-life examples in both lecture and lab.

Surface Frontal Analysis

In general, a surface front can be identified on a weather map utilizing the following guidelines:

- Start within a warm air mass and move toward where you believe the cold air to be located. The location(s) where the temperature begins to decrease rapidly over a short distance provide your first-guess estimate for the location of the frontal boundary.
- Repeat this exercise for moisture, moving from moister toward drier conditions, whether in the context of dewpoint temperature (T_d), mixing ratio (q or w), or another thermodynamic variable such as equivalent potential temperature (θ_e).
- Utilizing an analysis of sea level pressure, look for where the maximum troughing or "kinkiness" in the isobar field is located. The surface front is generally located within this region at the location(s) where sea level pressure is minimized.
- Utilizing a surface pressure tendency analysis, look for locations where the surface pressure is rapidly rising and rapidly falling in close proximity to one another. Surface fronts are typically found ahead of regions of rapidly rising sea level pressure and behind regions of rapidly falling sea level pressure.
- Utilizing an analysis of surface wind, look for locations where the wind direction changes rapidly (in a cyclonic manner) over a short distance. The surface front is generally located through such regions.
- The motion of surface fronts is fairly consistent through time. The placement of a surface front at a given time can often be approximated by extrapolating from a series of previous surface frontal analyses.
- Clouds and precipitation are generally found along surface frontal boundaries, as we will examine more closely shortly.
- Utilizing an analysis of temperature tendency, look for locations where the temperature is rapidly changing (either warming or cooling). Surface fronts are generally found on the leading edges of regions of rapid warming (warm fronts) or rapid cooling (cold fronts).

Cold Fronts

When a relatively cold air mass advances toward a relatively warm air mass, the boundary separating the two air masses is known as a cold front. Cold fronts are located along the leading edge of the cold air. In general, this is also the location where the wind direction rapidly changes and, thus, where relative vorticity is locally maximized. The cold frontal zone extends from the location of the cold front itself rearward to the point at which the temperature ceases to drop rapidly. This, admittedly, is somewhat of a subjective or qualitative criterion. These concepts are illustrated in Figures 1 and 2.

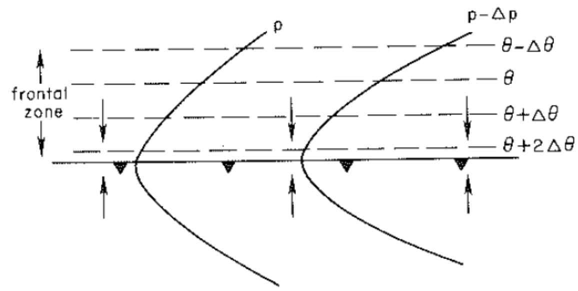

Figure 1. Surface isentropes (dashed lines), isobars (solid lines), and divergent wind (vectors) with an idealized surface cold front (triangled line) and cold frontal zone. Reproduced from *Synoptic-Dynamic Meteorology in Midlatitudes (Vol. II)* by H. Bluestein, their Figure 2.17.

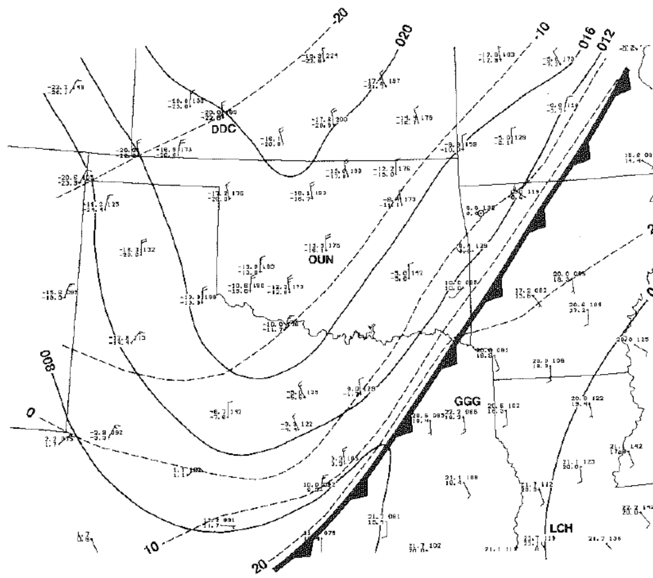

Figure 2. Surface analysis (station data), isobars (hPa; solid lines, leading 1 omitted), and isotherms (°C; dashed lines) ahead of a strong surface cold front (triangled line). Reproduced from *Synoptic-Dynamic Meteorology in Midlatitudes (Vol. II)* by H. Bluestein, their Figure 2.18.

A cold front is associated with strong vertical wind shear. Ahead of a cold front, the wind veers with increasing height, indicative of layer-mean warm air advection; behind a cold front, the wind backs with increasing height, indicative of layer-mean cold air advection. Within the relatively cold air, the lapse rate can approach dry adiabatic, implying the presence of turbulent vertical mixing and the potential for strong, gusty winds at the surface.

The actual cold frontal zone, or the region over which temperature and moisture change rapidly over a short horizontal distance, has a vertical depth of 500 m to 1500 m. Within this frontal zone, potential temperature rapidly increases with increasing altitude, a stable situation. Viewed on a sounding, a strong inversion in both the temperature and dew point temperature traces is noted; this is what is known as a *frontal inversion*. A representative example is given in Figure 3.

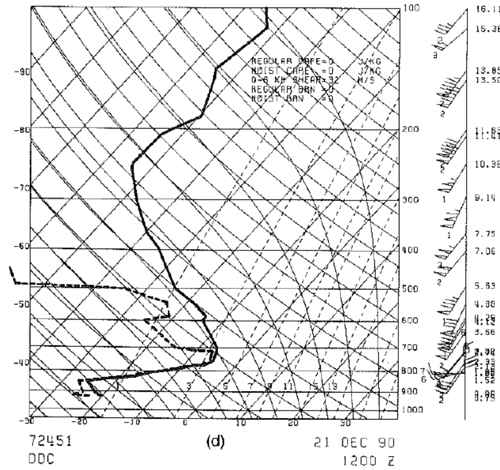

Figure 3. Sounding through a cold frontal zone, including temperature (solid black line), dew point temperature (dashed black line), and wind (barbs; half: 5 kt, full: 10 kt, pennant: 50 kt). Note the frontal inversion centered on 800 hPa. Reproduced from *Synoptic-Dynamic Meteorology in Midlatitudes (Vol. II)* by H. Bluestein, their Figure 2.19d.

Cold fronts slope upward in the rearward direction; for example, a cold front moving to the southeast is found further to the northwest at progressively higher altitudes. This is a very important distinguishing characteristic of mid-latitude cyclones! The typical vertical slope of a cold frontal zone is on the order of 1/100, such that it rises 1 km for every 100 km of horizontal distance. Its slope is greatest (~1/50) near the surface and smallest (~1/150 to ~1/200) at higher altitudes. A vertical cross-section through a representative cold front is illustrated in Figure 4.

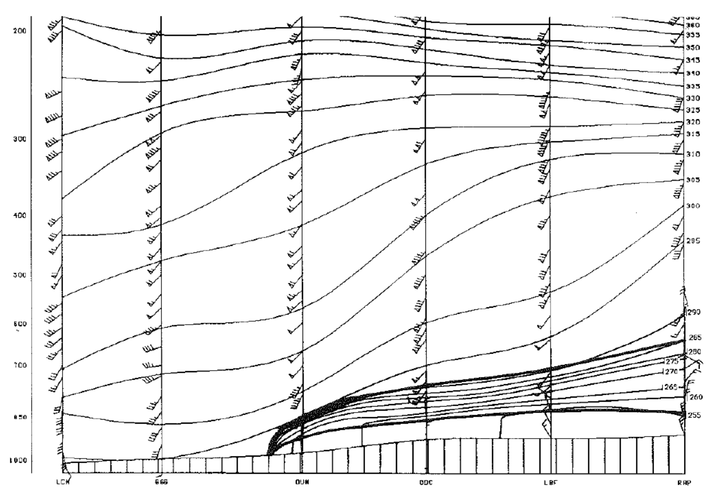

Figure 4. Vertical cross-section (south at left, north at right) of potential temperature (K; solid lines) and winds (barbs; half: 5 kt, full: 10 kt, pennant: 50 kt) through a cold frontal zone. The thick black lines in the lower right demarcate the rearward-sloping cold frontal zone. Reproduced from *Synoptic-Dynamic Meteorology in Midlatitudes (Vol. II)* by H. Bluestein, their Figure 2.20.

Cold frontal motion is determined by the magnitude of the wind behind the cold front (i.e., within the cold air) perpendicular to the cold front. Cold fronts move rapidly when the behind-front wind is strong and oriented perpendicular to the cold front; they move less rapidly or become stationary when the behind-front wind is weak and/or oriented parallel to the cold front.

The distribution of clouds and precipitation near the front depends upon the horizontal distributions of vertical motion, stability, the front-relative flow, and moisture. Clouds and precipitation are generally located where ascent, weaker stability or greater instability, and sufficient moisture are co-located. This, naturally, is generally in close proximity to a cold front.

A cold front that is characterized by clouds and precipitation primarily along and ahead of the cold front is known as a *katafront*; a cold front that is characterized by clouds and precipitation primarily along and behind the cold front is known as an *anafront*. Ascent that slopes backward with height over the front leads to an anafront structure. Conversely, ascent that is upright or slopes forward with height ahead of a front leads to a katafront structure.

Warm Fronts

When a relatively cold air mass retreats in advance of a relatively warm air mass, the boundary separating the two air masses is known as a warm front. Most, but not all, surface cyclones are associated with warm fronts. Warm fronts are located along the rear edge of the advancing warm air, as illustrated in Figure 5. The warm frontal zone extends from the location of the warm front itself forward (away) to the point at which the temperature ceases to decrease rapidly.

As with cold fronts, warm fronts are associated with strong vertical wind shear. Ahead of a warm front, the wind veers with increasing height, indicative of layer-mean warm air advection. The actual warm frontal zone has a vertical depth of 500 m to 1500 m over which potential temperature increases rapidly with increasing altitude, similar to cold fronts. However, warm fronts are generally not as strong, nor as intense, as cold fronts.

Warm fronts slope upward in the forward direction; for example, a warm front moving to the north is found further to the north at progressively higher altitudes. However, the vertical slope of a warm frontal zone is shallower than that of a cold frontal zone: greatest (~1/150) near the surface and smallest (~1/200 to 1/300) at higher altitudes. A vertical cross-section through a representative warm front is illustrated below in Figure 6.

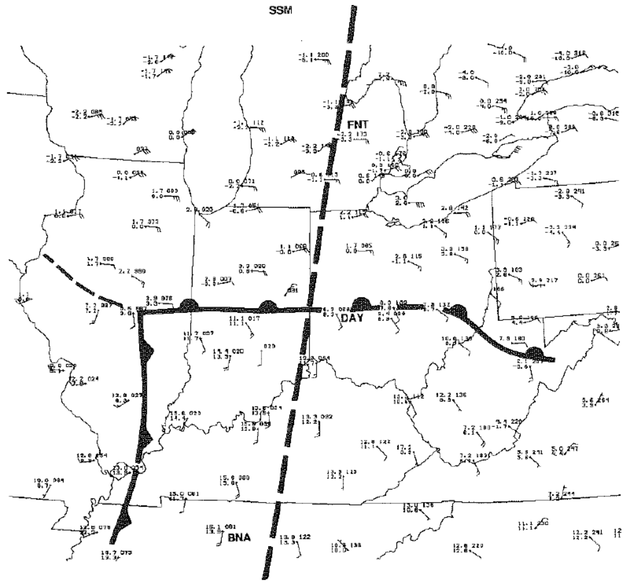

Figure 5. Surface analysis (station data) ahead of a surface warm front (ovaled line) and surface cold front (triangled line). Note the change in wind direction and temperature across the warm front. Reproduced from *Synoptic-Dynamic Meteorology in Midlatitudes (Vol. II)* by H. Bluestein, their Figure 2.26.

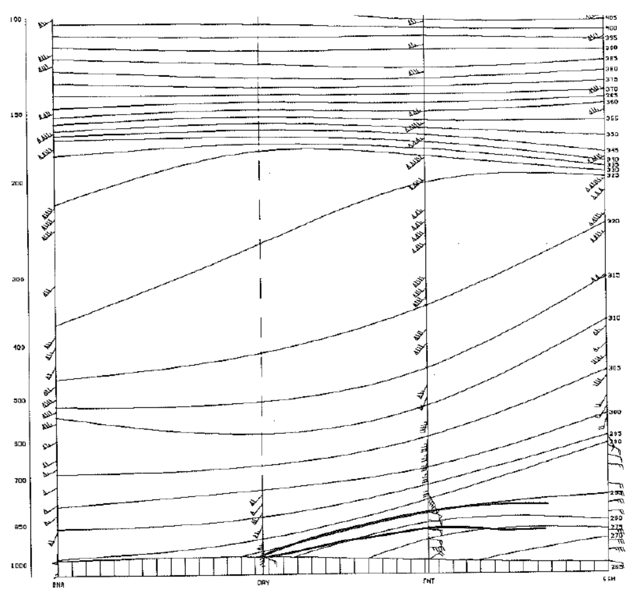

Figure 6. As in Figure 4, except for a warm frontal zone (thick black lines). Reproduced from *Synoptic-Dynamic Meteorology in Midlatitudes (Vol. II)* by H. Bluestein, their Figure 2.28.

Warm frontal motion is largely determined by the magnitude of the wind ahead of the warm front (i.e., within the cold air) perpendicular to the warm front itself. However, as the wind ahead of a warm front typically has a large along-front component, warm fronts typically do not move as rapidly as do cold fronts.

Stationary Fronts

When the synoptic-scale wind in the rear of a cold front or ahead of a warm front becomes oriented largely parallel to the front itself, the movement of the front slows. When it reaches a sufficiently small speed (generally ≤ 2.5 m s^{-1}), it is said to become stationary. In such cases, the colder air mass does not advance toward or retreat from the warmer air mass. Precipitation associated with stationary fronts is generally of a stratiform (i.e., non-convective) nature; however, stationary fronts often are foci for the development and movement of periodic mesoscale convective systems during the warm season(s).

Occluded Fronts and Cyclone Seclusion

When a cold front overtakes a warm front equatorward of a cyclone, the resulting feature is called an occlusion. The surface boundary where the cold front meets the warm front is called an occluded front. There are two types of occluded fronts: warm occlusions, in which the advancing cold air is warmer than the retreating cold air, and cold occlusions, in which the advancing cold air is colder than the retreating cold air. Both types of occlusions are illustrated in Figure 7.

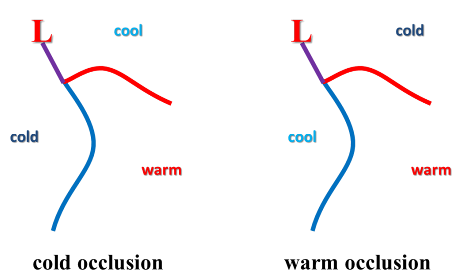

Figure 7. Idealized depiction of the air mass contrasts associated with cold occlusions (left) and warm occlusions (right).

It is difficult to find a true occlusion, whether warm or cold, in the real world. Nevertheless, the development of an occluded front represents the pinnacle of cyclone development, or maturity, as viewed from the traditional Bergen School frontal perspective. We will examine the life cycle of mid-latitude synoptic-scale cyclones in detail in a later lecture.

An occlusion can form on the cold air side of a deepening cyclone even if a cold front does not overtake a warm front. In such cases, relatively warm air becomes trapped in the immediate vicinity of the deepening cyclone that is surrounded by cold air. The warm air becomes entangled with the cyclone as a consequence of *frontal fracture*, where the warm and cold front become finitely disconnected from one another for a brief period of time.

The resultant cyclone structure is what is known as a *warm seclusion*, and warm seclusion extratropical cyclones are often among the most intense of all extratropical cyclones and are most common over maritime regions. The structure of a warm seclusion is illustrated in Figure 8. The development of a warm seclusion is presented as an alternative hallmark of the pinnacle of cyclone development, and the life cycle of a mid-latitude cyclone that leads to a warm seclusion will be examined in a subsequent lecture.

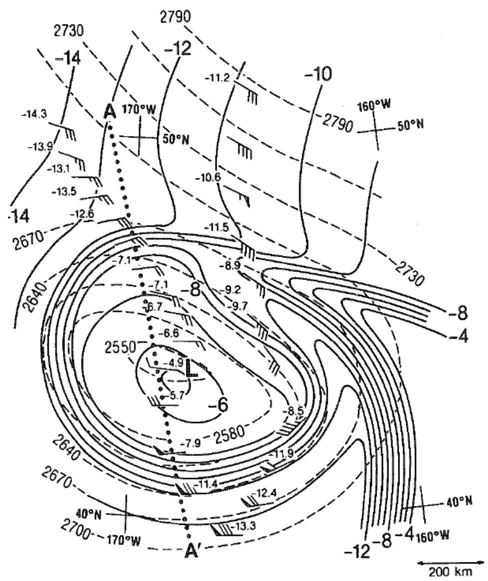

Figure 8. Analysis of surface isotherms (°C; solid lines), 700 hPa height (m; dashed lines), and aircraft-obtained winds (barbs; half: 5 kt, full: 10 kt, pennant: 50 kt) depicting a warm seclusion extratropical cyclone. Reproduced from *Synoptic-Dynamic Meteorology in Midlatitudes (Vol. II)* by H. Bluestein, their Figure 2.33.

Middle-to-Upper Tropospheric Fronts

An upper tropospheric front is a zone of strong, quasi-horizontal temperature gradient in the middle to upper troposphere. Such a front is typically accompanied by large static stability, similar to their surface or lower tropospheric counterparts. These fronts often form downstream of an upper tropospheric ridge and upstream from an upper tropospheric trough in a region of synoptic-scale descent and may propagate around the base of the downstream synoptic-scale trough. Concurrently, upper tropospheric vertical motions – both upward and downward – are thought to be important with upper tropospheric fronts, and these fronts are often associated with localized *tropopause folds*, or localized tropopause lowerings. A vertical cross-section through an upper tropospheric front is presented in Figure 9.

Upper tropospheric fronts are located along the leading edge of the quasi-horizontal temperature gradient aloft. However, temperature advection associated with these fronts is relatively weak and, as a result, we do not refer to them as either warm or cold fronts as we do for surface fronts. Likewise, there is often negligible sensible weather found in conjunction with these fronts. The presence of a quasi-horizontal temperature gradient, from thermal wind balance, indicates that upper tropospheric fronts are typically associated with upper-level jets and/or jet streaks. The strong vertical wind shear that often accompanies these fronts is often accompanied by turbulence, often of the clear-air variety, that can disrupt air travel.

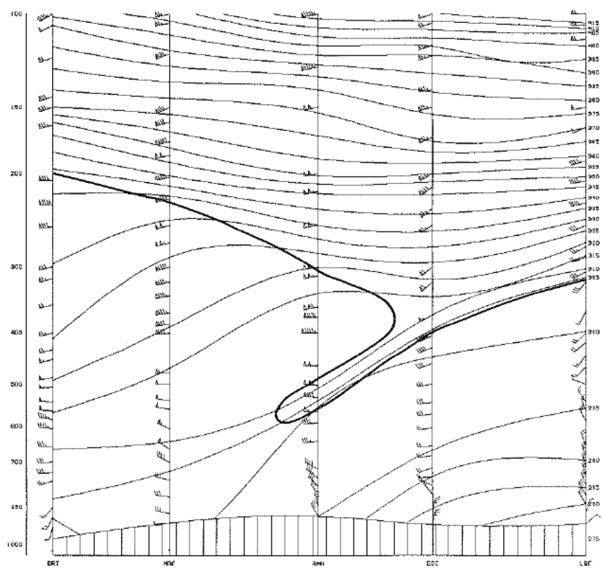

Figure 9. Vertical cross-section (south at left, north at right) through an upper tropospheric front. Potential temperature (K) is depicted by the solid contours every 5 K, while wind barbs (half: 5 kt, full: 10 kt, pennant: 50 kt) are depicted at five locations along the cross-section. The location of the tropopause is approximated by the thick black line. Note the tropopause fold between on the south side of the frontal zone. Reproduced from *Synoptic-Dynamic Meteorology in Midlatitudes (Vol. II)* by H. Bluestein, their Figure 2.47.

Other Surface and Near-Surface Frontal Circulations

Drylines

The dryline is a narrow zone of strong horizontal gradients of moisture near the surface. It is a climatological feature of the southern and central Great Plains of the United States, primarily during the warm season, and separates moist maritime air from the Gulf of Mexico to the east from dry continental air from the southwestern United States and Mexico to the west. As terrain slopes upward to the west, the vertical depth of the moist air is shallower to the west and deeper to the east. An idealized vertical cross-section through a dryline is depicted in Figure 10.

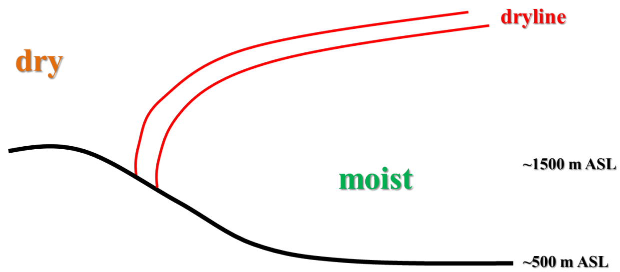

Figure 10. Idealized vertical cross-section, with west to the left and east to the right, through the dryline across the southern Great Plains.

Temperature contrasts across the dry line are often driven by sensible heating differences between the dry air to the west and moist air to the east, with warmer temperatures found to the west of the dryline during the day and to the east of the dryline at night. There is often a subtle wind shift along the dry line such that the dry line is a region of localized confluence. This provides a mesoscale source of forcing for ascent. An analysis of a dry line is presented in Figure 11. The dryline is a climatologically-favored location for thunderstorm development during the warm season.

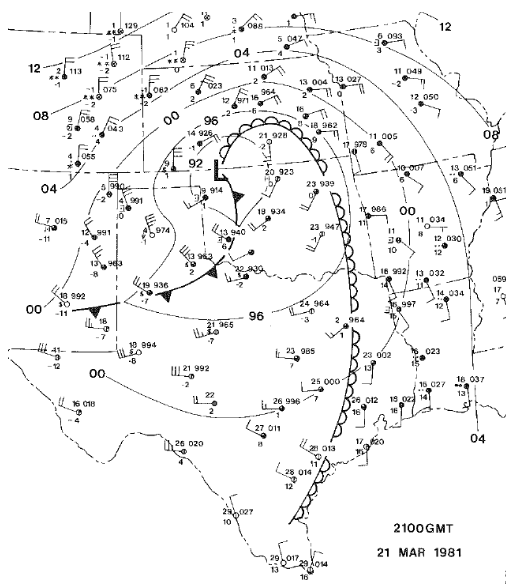

Figure 11. Surface analysis, depicting station observations (T and T_d in °C, $\boldsymbol{v}$ with a half barb = 5 kt and full barb = 10 kt) and isobars (hPa; contours with leading 9 or 10 omitted) across the south-central United States. The dryline is depicted by the open-scalloped line across Kansas, Oklahoma, and Texas. Reproduced from *Synoptic-Dynamic Meteorology in Midlatitudes (Vol. II)* by H. Bluestein, their Figure 2.40.

Dryline motion is generally to the east during the day and to the west during the night. Its motion during the day is controlled by turbulent vertical mixing as driven by sensible heating. Turbulent mixing acts to mix the dry air above with the moist air below. This happens most rapidly to the west, where moisture is shallower. Such mixing reduces the dewpoint temperature (and mixing ratio) at the surface, in effect “pushing” the dryline eastward. At night, turbulent vertical mixing ends as sensible heating (itself driven by insolation) ends. Radiative cooling of the boundary layer occurs and is more efficient in the drier air. This enables the dryline to “push” westward as the surface is radiatively cooled toward its dewpoint temperature.

Sea Breezes, Lake Breezes, and Land Breezes

Sea and lake breezes are two-dimensional (across-boundary, vertical) features oriented parallel to the coastline. Bodies of water of all sizes, from relatively small lakes to the great oceans of the world, can give rise to sea or lake breezes. A sea breeze and a lake breeze are identical to one another; the preferred term is often a matter of local choosing.

Sea and lake breezes result from differential daytime heating of a land mass as compared to that of an adjacent body of water. As it takes a greater amount of heat energy to warm a given mass of water by 1°C than it does to warm an identical mass of land by 1°C, daytime heating can lead to a situation where the land is relatively warm compared to the adjacent body of water. This is most common during the warm season.

Under such conditions, since the temperature of a vertical layer is directly proportional to the thickness of that layer, the near-surface thickness over land becomes greater than that over water. This results in locally low pressure at the surface over land, as the lower isobaric surface is depressed downward where thickness is high, and locally high pressure at the surface over water, as the lower isobaric surface is elevated upward where thickness is low. The resulting flow from high toward low pressure acts like a density current to advect relatively cool air from over the body of water to locations over land.

As with other frontal boundaries, there exists localized confluence along the leading edge of a sea or lake breeze. Such confluence extends upward from the surface to approximately 1-2 km above ground level, representing the vertical depth of the sea breeze circulation. (Above this height, the flow actually is directed from land to sea, as can also be illustrated using thickness-related arguments.) The propagation of a sea or lake breeze is controlled by the intensity of the land-water temperature contrast – the sea breeze front is stronger and faster-moving when the temperature contrast is larger – and by the strength and direction of the synoptic-scale flow.

At night, land cools faster than does water. As a result, the same sort of circulation between land and water – except in the opposite direction! – can develop. This is what is known as a land breeze. The localized confluence along land breezes can lead to nocturnal convection over the adjacent body of water if other factors favorable for convective development are present.

For Further Reading

Sections 6.1, 6.4, and 6.5 of *Midlatitude Synoptic Meteorology* by G. Lackmann cover characteristics of fronts, both at and above the surface, in great detail. Sections 2.1 and 2.4 of *Synoptic-Dynamic Meteorology in Midlatitudes, Vol. II* by H. Bluestein go into extensive detail regarding the structure of fronts. Fronts and air masses are discussed in Chapters 8 and 9, respectively, of *Weather Analysis* by D. Djurić. Much of this lecture has been derived from these two references. Basic information regarding air masses and the structure of cold and warm fronts can be found in nearly any introductory meteorology textbook, including Chapter 9 of *Meteorology* (4th ed.) by S. Ackerman and J. Knox.