<u>Synoptic Meteorology I: Fronts</u>

For Further Reading

Sections 6.1, 6.4, and 6.5 of *Midlatitude Synoptic Meteorology* by G. Lackmann describe frontal characteristics, both at and above the surface, in detail. Sections 2.1 and 2.4 of *Synoptic-Dynamic Meteorology in Midlatitudes, Vol. II* by H. Bluestein describe frontal structure in exhaustive detail. Fronts and air masses are discussed in Chapters 8-9 of *Weather Analysis* by D. Djurić.

Introduction

In the atmosphere, we typically observe large temperature gradients and strong winds concentrated in localized areas. The former are associated with fronts, while the latter are associated with jets. Why are fronts and jets typically linked and located close to each another?

Thermal wind balance, which relates the magnitude and direction of the layer-mean horizontal temperature gradient (which is often large near a front) to a measure of vertical wind shear (which is often strong in the presence of a jet), helps us understand why fronts and jets are often found near one another. However, it does not explain how 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, despite the finite lines that are often drawn on weather maps to depict them) of locally large horizontal temperature gradients. What do we mean by *elongated* and *locally large*, 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 large**: 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 value change per 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. For a cold front, this inversion separates the post-frontal near-surface cold air mass from a warmer pre-frontal air mass underneath which the surface cold air mass has surged forward. For a warm front, this inversion separates the pre-frontal near-surface cold air mass from a warmer post-frontal air mass that has ascended over the surface warm front. We will document these structures in more detail shortly.

2) *What is a jet?*

A jet is an intense, narrow, quasi-horizontal current of wind that is associated with large 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).
- **Large**: 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 wind speed maximum embedded in 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 like precipitation, thunderstorms, strong winds, intense cyclone development, etc.
- Large 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 particulate transport, such as ozone and/or pollutants.

This lecture focuses on fronts, whereas the next lecture focuses on jets and jet streaks.

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. 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 and are generally located downstream (ahead) of upper-tropospheric troughs and upstream of (behind) upper-tropospheric ridges in a synoptic-scale region of ascent.

Surface fronts are characterized by one or more of the following properties:

- A zone of large, across-front temperature, moisture, vertical motion, and relative vorticity (horizontal rotation) 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 (horizontal wind directed *toward* a single axis) 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.

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 features 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.

Cold Fronts

When a colder air mass *advances toward* a warmer 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 at the surface. 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 Figs. 1 and 2.

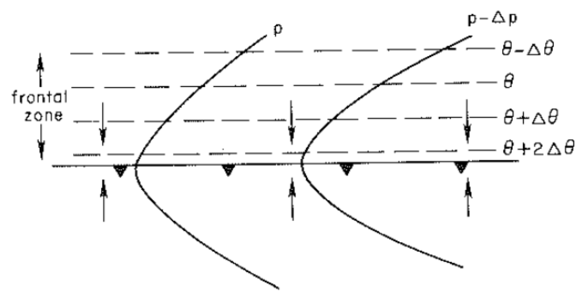

Figure 1. Surface isentropes (dashed lines), isobars (solid lines), and divergent wind (vectors) with an idealized surface cold front (triangled line, with the triangles pointing in the direction of cold front motion) and frontal zone. Reproduced from *Synoptic-Dynamic Meteorology in Midlatitudes (Vol.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

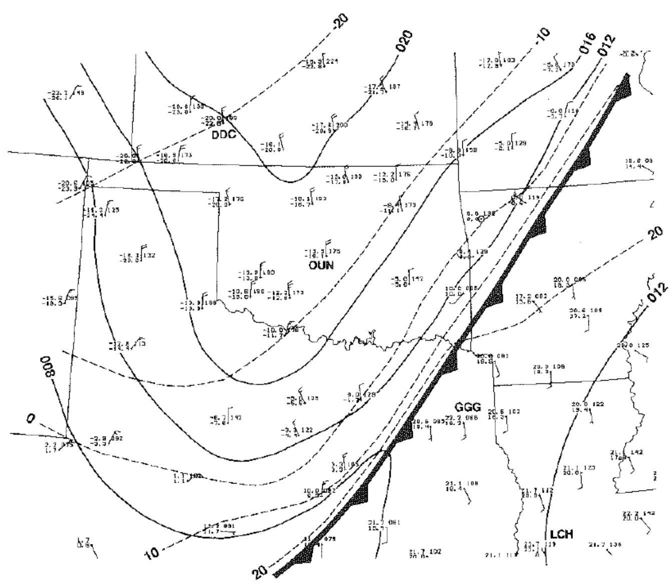

Figure 2. Surface analysis (station data), isobars (hPa; solid lines, leading 1 omitted), and isotherms (°C; dashed lines) ahead of a strong surface cold front (triangled line). Reproduced from *Synoptic-Dynamic Meteorology in Midlatitudes (Vol. II)* by H. Bluestein, their Fig. 2.18.

A cold front is associated with large 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 between the surface and the bottom of the frontal zone can approach dry adiabatic, implying the presence of turbulent vertical mixing and the potential for strong and gusty surface winds.

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. Potential temperature rapidly increases with increasing altitude from the frontal zone upward to the pre-frontal air mass that resides atop it. Viewed on a sounding, a sharp inversion in both the temperature and dew point temperature traces is noted; this is what is known as a *frontal inversion* and represents a situation that is stable to upward parcel displacements. A representative example is given in Fig. 3.

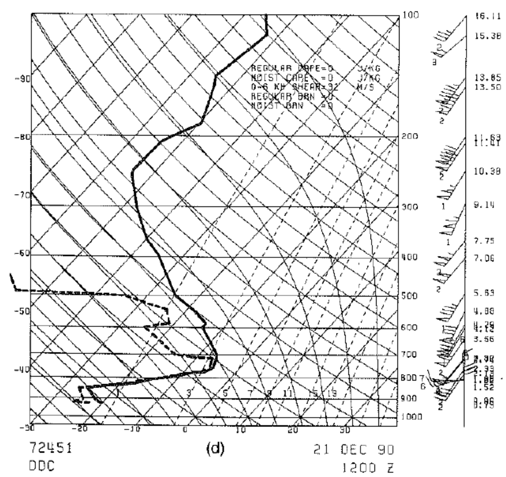

Figure 3. Sounding through a cold frontal zone, including temperature (solid black line), dew point temperature (dashed black line), and wind (barbs; half: 5 kt, full: 10 kt, pennant: 50 kt). Note the frontal inversion centered on 800 hPa. Reproduced from *Synoptic-Dynamic Meteorology in Midlatitudes (Vol. II)* by H. Bluestein, their Fig. 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 midlatitude 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 Fig. 4.

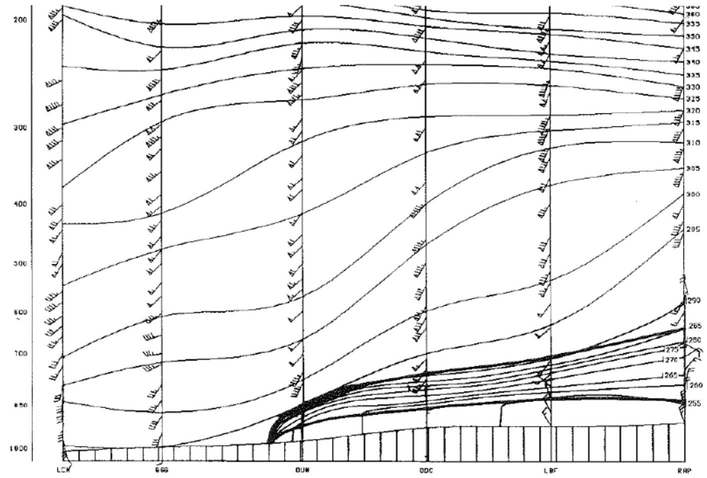

Figure 4. Vertical cross-section (south at left, north at right) of potential temperature (K; solid lines) and winds (barbs; half: 5 kt, full: 10 kt, pennant: 50 kt) through a cold frontal zone. The thick black lines in the lower right demarcate the rearward-sloping cold frontal zone. Note that we typically use potential temperature, rather than temperature, to diagnose frontal zones on vertical cross-sections. Reproduced from *Synoptic-Dynamic Meteorology in Midlatitudes (Vol. II)* by H. Bluestein, their Fig. 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, as is often found near the leading edge of a surface cold front. Cold fronts characterized by clouds and precipitation primarily along and ahead of the cold front are known as *katafronts*; cold fronts characterized by clouds and precipitation primarily along and behind the cold front are known as *anafronts*.

Katafronts are characterized by ascent in the pre-frontal warm air mass, rather than along or behind the surface cold front. Katafronts are typically accompanied by cooler, drier air aloft pushing ahead of the surface cold front, resulting in relatively warm and moist air near the surface being located beneath relatively cool and dry middle tropospheric air. The associated large change in temperature with increasing height increases the likelihood that near-surface air will be able to rise over a great vertical distance to become saturated and form pre-frontal precipitation as it does so. Conversely, anafronts are characterized by comparatively weak ascent over the surface cold front, resulting in precipitation of light to moderate intensity in the cold air. Katafront and anafront schematics are provided in Figs. 6.18b and 6.20b, respectively, of *Midlatitude Synoptic Meteorology*.

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 Fig. 5. The warm frontal zone extends from the location of the warm front itself forward to the point at which the temperature ceases to decrease rapidly.

As with cold fronts, warm fronts are associated with large vertical wind shear. Ahead of a warm front, the wind veers with increasing height, indicating layer-mean warm air advection. The actual warm frontal zone has a vertical depth of 500-1500 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, and thus the increase in potential temperature with increasing altitude across a warm frontal zone is not as large as is seen with 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 provided by Fig. 6.

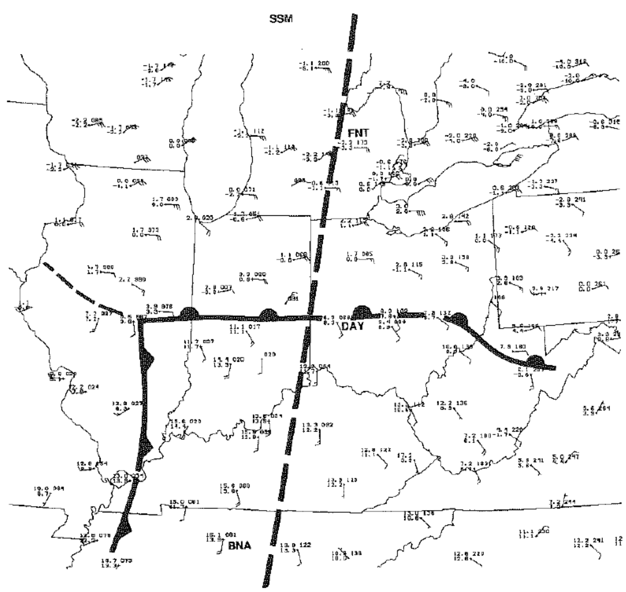

Figure 5. Surface analysis (station data) ahead of a surface warm front (ovaled line, with the ovals pointing in the direction of warm front motion) and surface cold front (triangled line). Note the change in wind direction and temperature across the warm front. Reproduced from *Synoptic-Dynamic Meteorology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

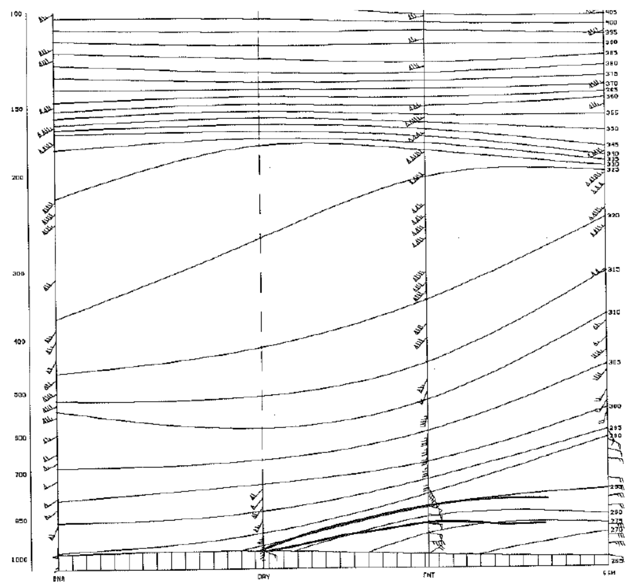

Figure 6. As in Fig. 4, except for a warm frontal zone (thick black lines). Reproduced from *Synoptic-Dynamic Meteorology in Midlatitudes (Vol. II)* by H. Bluestein, their Fig. 2.28.

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 stratiform (or non-convective; i.e., light to moderate in intensity in association with primary stratus clouds) nature; however, stationary fronts often are foci for the development and movement of periodic mesoscale convective systems during spring and summer. Stationary fronts are denoted on surface charts by alternating blue-triangled and red-ovaled lines, with the triangles pointing toward the warm air and the ovals pointing toward the cold air.

Occluded Fronts and Cyclone Seclusion

When a warm air mass ahead of a cold front begins to become horizontally elongated and wrap up diffusively around a cyclone, resulting in the cyclone's warm and cold fronts merging, the resulting frontal structure is known as an *occluded front*. Occluded fronts are typically denoted by purple lines on surface charts, with alternating ovals and triangles pointing outward from the cyclone.

An occluded front typically separates air that was originally behind a cold front from air that was originally ahead of a warm front. In other words, an occluded typically separates two initially cold air masses. Occluded fronts slope upward and over whichever of these air masses that has greater static stability (i.e., a greater increase in potential temperature with height). Since the air behind a cold front tends to be more well-mixed than that ahead of a warm front, and consequently potential temperature increases more rapidly in the air mass initially ahead of a warm front, occluded fronts frequently slope upward and over the air mass initially found ahead of a warm front – sometimes referred to as a warm-type occlusion.

Occluded front development represents the pinnacle of cyclone development, or maturity, in the Bergen School model of a midlatitude cyclone's lifecycle. In this model, an occluded front forms as its attendant cyclone reaches its lifetime maximum intensity. However, many cyclones continue to deepen – sometimes substantially (10-30 hPa) following occluded front development. We will consider the Bergen School midlatitude cyclone lifecycle in detail at the start of next semester.

An occlusion can also form on the cold side of a deepening cyclone even if its cold front does not overtake its warm front. In such cases, relatively warm air becomes isolated near the center of the deepening cyclone, with relatively cold air surrounding the localized warm air. Warm air becomes entangled with the cyclone due to *frontal fracture*, where the warm and cold front become finitely disconnected from one another for a short 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 Fig. 7. The development of a warm seclusion is as an alternate hallmark of the pinnacle of cyclone development, and the life cycle of a midlatitude cyclone that leads to a warm seclusion will be examined in a subsequent lecture.

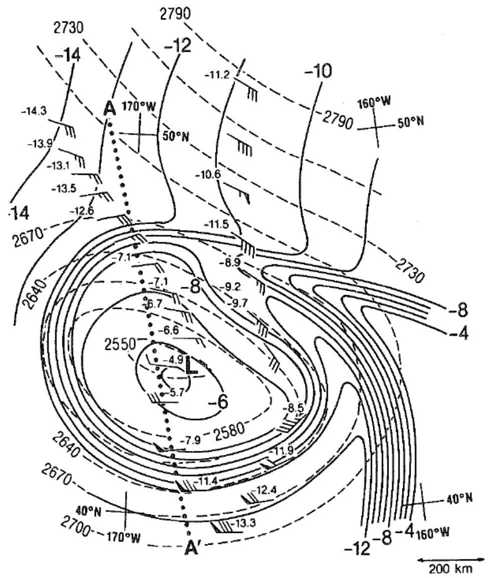

Figure 7. Analysis of surface isotherms (°C; solid lines), 700 hPa height (m; dashed lines), and aircraft-obtained winds (barbs; half: 5 kt, full: 10 kt, pennant: 50 kt) depicting a warm seclusion extratropical cyclone. Reproduced from *Synoptic-Dynamic Meteorology in Midlatitudes (Vol. II)* by H. Bluestein, their Fig. 2.33.

Middle- and 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 Fig. 8.

Upper tropospheric fronts are located along the leading edge of the quasi-horizontal temperature gradient aloft. However, 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 large vertical wind shear that often accompanies these fronts is often accompanied by turbulence, often of the clear-air variety, that can disrupt air travel.

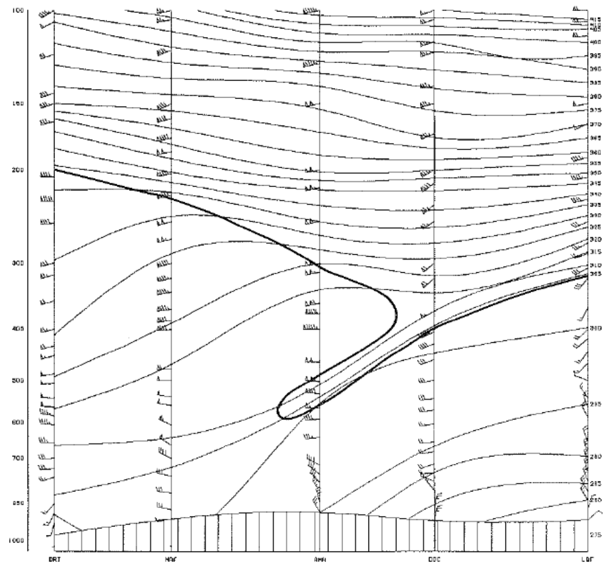

Figure 8. Vertical cross-section (south at left, north at right) through an upper-tropospheric front. Potential temperature (K) is depicted by the solid contours every 5 K, while wind barbs (half: 5 kt, full: 10 kt, pennant: 50 kt) are depicted at five locations along the cross-section. The location 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 Fig. 2.47.

Other Surface and Near-Surface Frontal Circulations

Drylines

A dryline is a narrow zone of large near-surface horizontal moisture gradients. It is a climatological feature of the southern and central Great Plains of the United States, primarily during the warm season, and separates mT air from the Gulf of Mexico to the east from cT from the southwestern United States and Mexico to the west. As terrain slopes upward and the distance from the Gulf of Mexico increases 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 Fig. 9.

Temperature contrasts across the dryline result from 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 (where more solar radiation can go into sensible heating than to evaporating surface water) during the day and to the east of the dryline (where the moister air less readily allows for radiative cooling) at night. There is often a subtle wind shift along the dryline, such that the dryline is a region of localized confluence and mesoscale ascent. An analysis of a dry line is presented in Fig. 10. The dryline is a climatologically favored location for thunderstorm development in spring and summer.

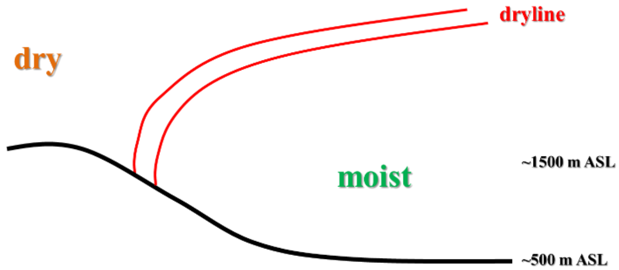

Figure 9. Idealized vertical cross-section, with west to the left and east to the right, through the dryline across the southern Great Plains.

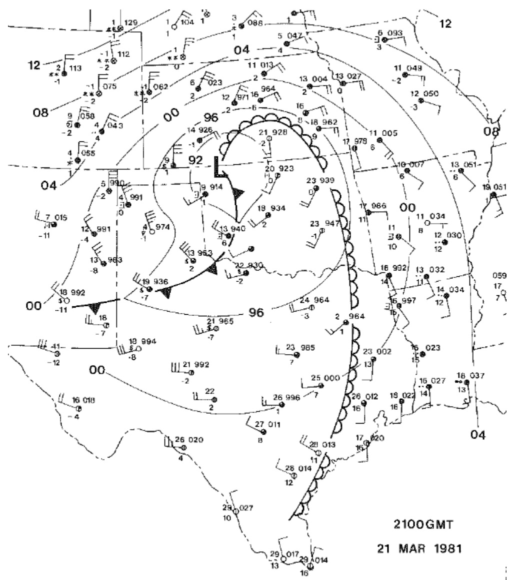

Figure 10. Surface analysis, depicting station observations (T and T_d in °C, $\boldsymbol{v}$ with a half barb = 5 kt and full barb = 10 kt) and isobars (hPa; contours with leading 9 or 10 omitted) across the south-central United States. The dryline is depicted by the open-scalloped line across Kansas, Oklahoma, and Texas. Reproduced from *Synoptic-Dynamic Meteorology in Midlatitudes (Vol. II)* by H. Bluestein, their Fig. 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 the turbulent vertical mixing that results from sensible heating. Turbulent mixing mixes dry air above with moist air below. This happens most rapidly to the west, where the near-surface moist layer is shallower (and turbulent eddies can be small in depth to reach the drier air above). Turbulent mixing reduces the surface dew point temperature and mixing ratio, in effect mixing the dryline eastward. At night, turbulent vertical mixing ends as sensible heating (resulting from insolation) terminates. Radiative cooling of the boundary layer occurs and is more efficient in the drier air. This enables the dryline to retreat or “push” westward as the surface is radiatively cooled toward its dew point 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 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.