12 FRONTS & AIRMASSES

Contents

A high-pressure center, or **high** (H), often contains an **airmass** of well-defined characteristics, such as cold temperatures and low humidity. When different airmasses finally move and interact, their mutual border is called a **front**, named by analogy to the battle fronts of World War I.

Fronts are usually associated with low-pressure centers, or **lows** (L, covered in the next chapter). Two fronts per low are most common, although zero to four are also observed. In the Northern Hemisphere, these fronts often rotate counterclockwise around the low center like the spokes of a wheel (Fig. 12.1), while the low moves and evolves. Fronts are often the foci of clouds, low pressure, and precipitation.

In this chapter you will learn the characteristics of anticyclones (highs). You will see how anticyclones are favored locations for airmass formation. Covered next are fronts in the bottom, middle, and top of the troposphere. Factors that cause fronts to form and strengthen are presented. This chapter ends with a special type of front called a dry line.

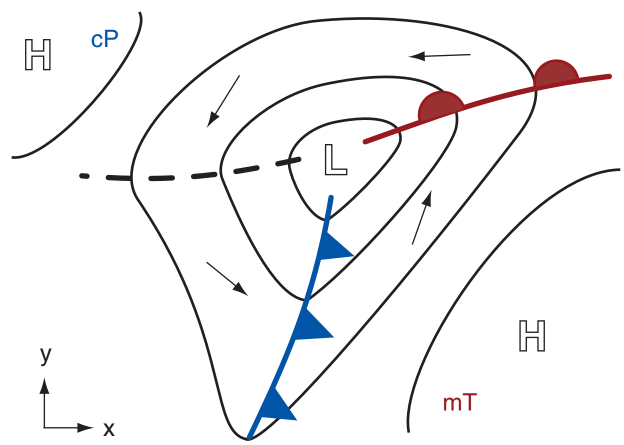


Figure 12.1
Idealized surface weather map (from the Weather Reports & Map Analysis chapter) for the N. Hemisphere showing high (H) and low (L) pressure centers, isobars (thin lines), a warm front (heavy red solid line with semicircles on one side), a cold front (heavy blue solid line with triangles on one side), and a trough of low pressure (black dashed line). Vectors indicate near-surface wind. cP indicates a continental polar airmass; mT indicates a maritime tropical airmass.

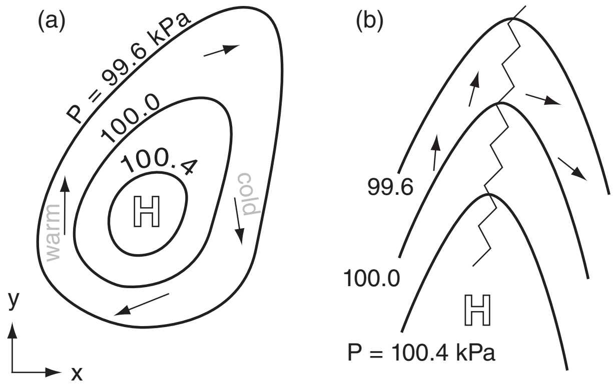


Figure 12.2
Examples of isobars plotted on a sea-level pressure map. (a) High-pressure center. (b) High-pressure ridge in N. Hemisphere mid-latitudes. Vectors show surface wind directions.

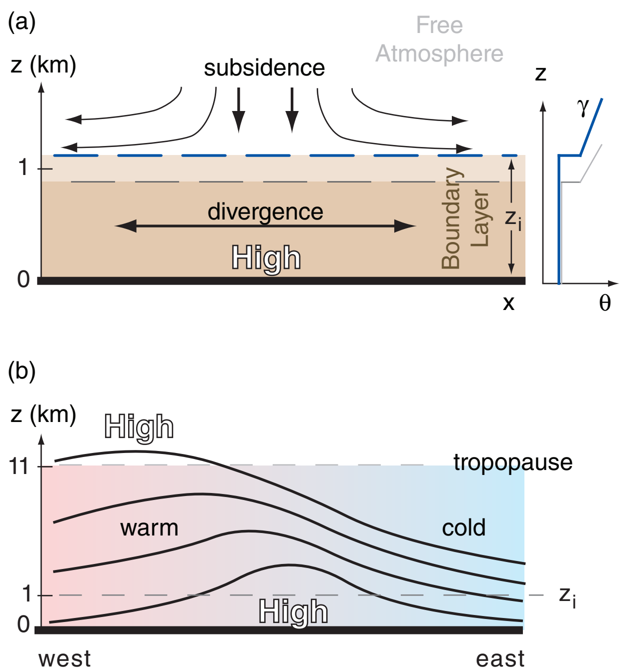


Figure 12.3
(a) Left: vertical circulation at a surface high-pressure center in the bottom half of the troposphere. Dark-blue dashed line marks the initial capping inversion at the top of the boundary layer. Grey dashed line shows the top later, assuming no turbulent entrainment into the boundary layer. Right: idealized profile of potential temperature, θ, initially (dark-blue line) and later (grey). The boundary-layer depth z_i is on the order of 1 km, and the potential-temperature gradient above the boundary layer is represented by γ.
(b) Tilt of high-pressure ridge westward with height, toward the warmer air. Curved lines are height contours of isobaric surfaces. Ridge amplitude is exaggerated in this illustration.

12.1. ANTICYCLONES OR HIGHS

12.1.1. Characteristics & Formation

High-pressure centers, or **highs**, are identified on constant altitude (e.g., sea-level) weather maps as regions of relative maxima in **pressure**. The location of high-pressure center is labeled with "H" (Fig. 12.2a). High centers can also be found on upper-air isobaric charts as relative maxima in **geopotential height** (see the Atmos. Forces & Winds chapter, Fig. 10.2).

When the pressure field has a relative maximum in only one direction, such as east-west, but has a horizontal pressure-gradient in the other direction, this is called a high-pressure **ridge** (Fig. 12.2b). The **ridge axis** is labeled with a zigzag line.

The column of air above the high center contains more air molecules than neighboring columns. This causes more weight due to gravity (see Chapter 1), which is expressed in a fluid as more pressure.

Above a high center is often downward motion (**subsidence**) in the mid-troposphere, and horizontal spreading of air (**divergence**) near the surface (Fig. 12.3a). Subsidence impedes cloud development, leading to generally clear skies and fair weather. Winds are also generally calm or light in highs, because gradient-wind dynamics of highs require weak pressure gradients near the high center (see the Atmos. Forces & Winds chapter).

The diverging air near the surface spirals outward due to the weak pressure-gradient force. Coriolis force causes it to rotate clockwise (anticyclonically) around the high-pressure center in the Northern Hemisphere (Fig. 12.2a), and opposite in the Southern Hemisphere. For this reason, high-pressure centers are called **anticyclones**.

Downward advection of dry air from the upper troposphere creates dry conditions just above the boundary layer. Subsidence also advects warmer potential temperatures from higher in the troposphere. This strengthens the temperature inversion that caps the boundary layer, and acts to trap pollutants and reduce visibility near the ground.

Subsiding air cannot push through the capping inversion, and therefore does not inject free-atmosphere air directly into the boundary layer. Instead, the whole boundary layer becomes thinner as the top is pushed down by subsidence (Fig. 12.3a). This can be partly counteracted by entrainment of free atmosphere air if the boundary layer is turbulent, such as for a convective mixed layer during daytime over land. However, the entrainment rate is controlled by turbulence in the boundary layer (see the Atmos. Boundary Layer chapter), not by subsidence.

Five mechanisms support the formation of highs at the Earth's surface:

• Global Circulation: Planetary-scale, semi-permanent highs predominate at 30° and 90° latitudes, where the global circulation has downward motion (see the General Circulation chapter). The **subtropical highs** centered near 30° North and South latitudes are 1000-km-wide belts that encircle the Earth. **Polar highs** cover the Arctic and Antarctic. These highs are driven by the global circulation that is responding to differential heating of the Earth. Although these highs exist year round, their locations shift slightly with season.

• Monsoons: Quasi-stationary, continental-scale highs form over cool oceans in summer and cold continents in winter (see the General Circulation chapter). They are seasonal (i.e., last for several months), and form due to the temperature contrast between land and ocean.

• Transient Rossby waves: Surface highs form at mid-latitudes, east of high-pressure ridges in the jet stream, and are an important part of mid-latitude weather variability (see the General Circulation and Extratropical Cyclone chapters). They often exist for several days.

• Thunderstorms: Downdrafts from thunderstorms (see the Thunderstorm chapters) create **meso-highs** roughly 10 to 20 km in diameter at the surface. These might exist for minutes to hours.

• Topography/Surface-Characteristics: Meso-highs can also form in mountains due to blocking or channeling of the wind, mountain waves, and thermal effects (anabatic or katabatic winds) in the mountains. Sea-breezes or lake breezes can also create meso-highs in parts of their circulation. (See the Regional Winds chapter.)

The actual pressure pattern at any location and time is a superposition of all these phenomena.

12.1.2. Vertical Structure

The location difference between surface and upper-tropospheric highs (Fig. 12.3b) can be explained using gradient-wind and thickness concepts.

Because of barotropic and baroclinic instability, the jet stream meanders north and south, creating troughs of low pressure and ridges of high pressure, as discussed in the General Circulation chapter. **Gradient winds** blow faster around ridges and slower around troughs, assuming identical pressure gradients. The region east of a ridge and west of a trough has fast-moving air entering from the west, but slower air leaving to the east. Thus, horizontal convergence of air at the top of the troposphere adds more air molecules to the whole tropospheric column at that location, causing a surface high to form east of the upper-level ridge.

West of surface highs, the anticyclonic circulation advects warm air from the equator toward the poles (Figs. 12.2a & 12.3b). This heating west of the surface high causes the **thickness** between isobaric surfaces to increase, as explained by the hypsometric equation. Isobaric surfaces near the top of the troposphere are thus lifted to the west of the surface high. These high heights correspond to high pressure aloft; namely, the upper-level ridge is west of the surface high.

The net result is that high-pressure regions tilt westward with increasing height (Fig. 12.3b). In the Extratropical Cyclone chapter you will see that deepening low-pressure regions also tilt westward with increasing height, at mid-latitudes. Thus, the mid-latitude tropospheric pressure pattern has a consistent phase shift toward the west as altitude increases.

12.2. AIRMASSES

An **airmass** is a widespread (of order 1000 km wide) body of air in the bottom third of the troposphere that has somewhat-uniform characteristics. These characteristics can include one or more of: temperature, humidity, visibility, odor, pollen concentration, dust concentration, pollutant concentration, radioactivity, cloud condensation nuclei (CCN) activity, cloudiness, static stability, and turbulence.

Airmasses are usually classified by their temperature and humidity, as associated with their source regions. These are usually abbreviated with a two-letter code. The first letter, in lowercase, describes the humidity source. The second letter, in upper-

Sample Application
A "**cA**" airmass has what characteristics?

Find the Answer
Given: cA airmass.
Find: characteristics

Use Table 12-1: cA = continental Arctic
Characteristics: **Dry and very cold**.

Check: Agrees with Fig. 12.4.
Exposition: Forms over land in the arctic, under the polar high. In Great Britain, the same airmass is labeled as Ac.

Table 12-1. Airmass abbreviations. **Boldface** indicates the most common ones.

Abbr.	Name	Description
c	**continental**	Dry. Formed over land.
m	**maritime**	Humid. Formed over ocean.
A	Arctic	Very cold. Formed in the polar high.
E	Equatorial	Hot. Formed near equator.
M	Monsoon	Similar to tropical.
P	**Polar**	Cold. Formed in subpolar area.
S	Superior	A warm dry airmass having its origin aloft.
T	**Tropical**	Warm. Formed in the subtropical high belt.
k	colder than the underlying surface	
w	warmer than the underlying surface	
Special (regional) abbreviations.		
AA	Antarctic	Exceptionally cold and dry.
r	returning	As in "rPm" returning Polar maritime [Great Britain].

Note: Layered airmasses are written like a fraction, with the airmass aloft written above a horizontal line and the surface airmass written below. For example, just east of a dryline you might have:

$$\frac{\text{cT}}{\text{mT}}$$

case, describes the temperature source. Table 12-1 shows airmass codes. [*CAUTION: In Great Britain, the two letters are reversed*.]

Examples are **maritime Tropical** (**mT**) airmasses, such as can form over the Gulf of Mexico, and **continental Polar** (**cP**) air, such as can form in winter over Canada.

After the weather pattern changes and the airmass is blown away from its genesis region, it flows over surfaces with different relative temperatures. Some organizations append a third letter to the end of the airmass code, indicating whether the moving airmass is (**w**) warmer or (**k**) colder than the underlying surface. This coding helps indicate the likely static stability of the air and the associated weather. For example, "mPk" is humid cold air moving over warmer ground, which would likely be statically unstable and have convective clouds and showers.

12.2.1. Creation

An airmass can form when air remains stagnant over a surface for sufficient duration to take on characteristics similar to that surface. Also, an airmass can form in moving air if the surface over which it moves has uniform characteristics over a large area. Surface high-pressure centers favor the formation of airmasses because the calm or light winds allow long residence times. Thus, many of the **airmass genesis** (formation) regions (Fig. 12.4) correspond to the planetary- and continental-scale high-pressure regions described in the previous section.

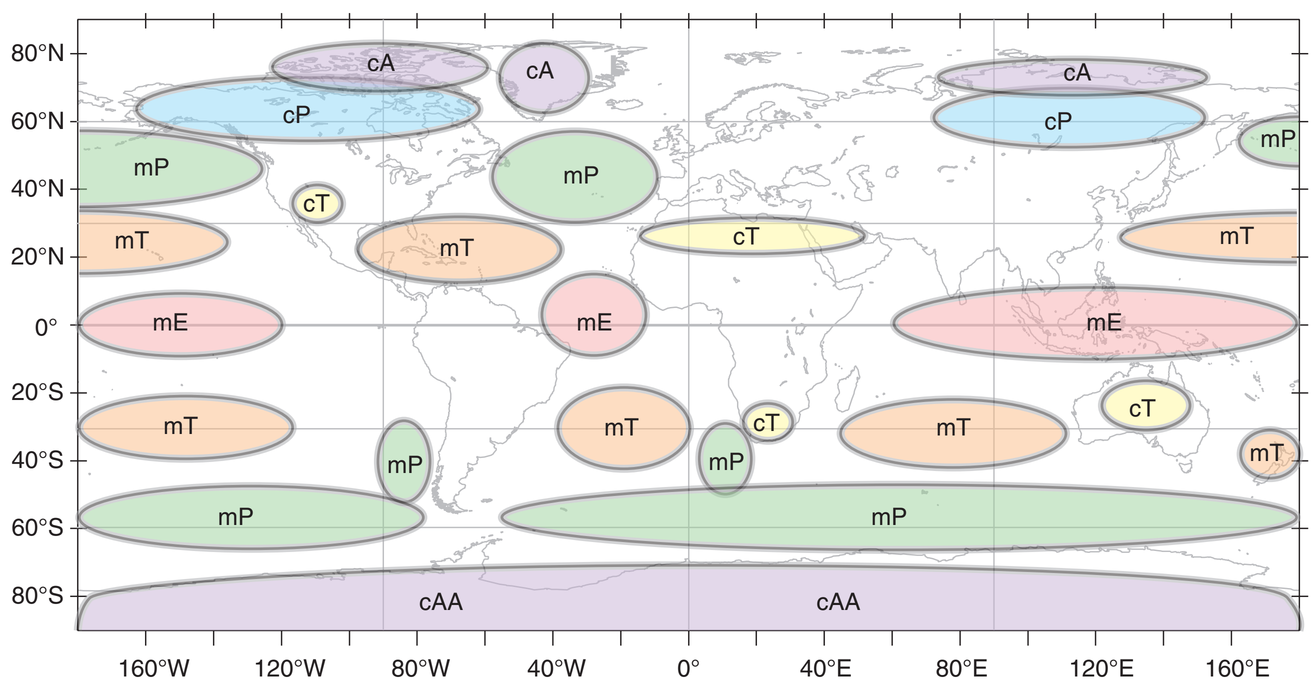


Figure 12.4
Airmass formation regions (symbols are defined in Table 12-1). Some meteorologists change all "mE" to "mT", and "cAA" to "cA".

Airmasses form as boundary layers. During their residence over a surface, the air is modified by processes including radiation, conduction, divergence, and turbulent transport between the ground and the air.

12.2.1.1. Warm Airmass Genesis

When cool air moves over a warmer surface, the warm surface modifies the bottom of the air to create an evolving, **convective mixed layer** (**ML**). Turbulence — driven by the potential temperature difference $\Delta\theta_s$ between the warm surface θ_{sfc} and the cooler airmass θ_{ML} — causes the ML depth z_i to initially increase (Fig. 12.5). This is the depth of the new airmass. A heat flux from the warm surface into the air causes θ_{ML} to warm toward θ_{sfc}. θ_{ML} is the temperature of the new airmass as it warms.

Synoptic-scale divergence β and subsidence w_s, which is expected in high-pressure airmass-genesis regions, oppose the ML growth. Changes within the new airmass are rapid at first. But as airmass temperature gradually approaches surface temperature, the turbulence diminishes and so does the rate of ML depth increase. Eventually, the ML depth begins to decrease (Fig. 12.5) because the reduced turbulence (trying to increase the ML thickness) cannot counteract the relentless subsidence.

A "toy model" describing the atmospheric boundary-layer processes that create a warm airmass is given in the INFO Box. The nearby Sample Application box uses this toy model to find the evolution of the warm airmass depth (i.e., the ML depth z_i) and its potential temperature θ_{ML} evolution. This is the solution that was plotted in Fig. 12.5.

The e-folding time (see Chapter 1) for the θ_{ML} to approach θ_{sfc} is surprisingly constant — about 1 to 2 days. As a result, creation of this warm **tropical** airmass is nearly complete after about a week. That is how long it takes until the airmass temperature nearly equals the surface temperature (Fig. 12.5). The time τ to reach the peak ML thickness is typically about 1 to 4 days (see another INFO box).

12.2.1.2. Cold Airmass Genesis

When air moves over a colder surface such as arctic ice, the bottom of the air first cools by conduction, radiation, and turbulent transfer with the ground. Turbulence intensity then decreases within the increasingly statically-stable boundary layer, reducing the turbulent heat transport to the cold surface.

However, direct radiative cooling of the air, both upward to space and downward to the cold ice surface, chills the air at a rate of 2°C day^{-1} (averaged over a 1 km thick boundary layer). As the air cools below the dew point, water-droplet clouds form. Continued radiative cooling from cloud top allows

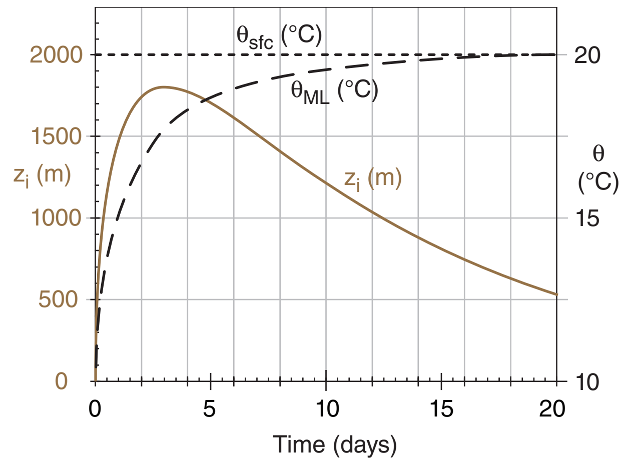


Figure 12.5
Genesis of a warm airmass after cold air comes to rest over a warmer surface. This is an example based on toy-model equations in the INFO Box. Airmass potential temperature is θ_{ML} and depth is z_i. Imposed conditions for this case-study example are: large-scale divergence $\beta = 10^{-6}$ s^{-1}, potential temperature gradient in vertical $\gamma_o = 3.3$ K km^{-1}, initial near-surface air potential temperature $\theta_{ML} = \theta_o = 10$°C, and surface temperature $\theta_{sfc} = 20$°C.

A SCIENTIFIC PERSPECTIVE • Math Clarity

In math classes, you might have learned how to combine many small equations into a single large equation that you can solve. For meteorology, although we could make such large single-equation combinations, we usually cannot solve them.

So there is no point in combining all the equations. Instead, it is easier to see the physics involved by keeping separate equations for each physical process. An example is the toy model given in the INFO Box on the next page for warm airmass genesis. Even though the many equations are coupled, it is clearer to keep them separate.

INFO • Bergen School of Meteorology

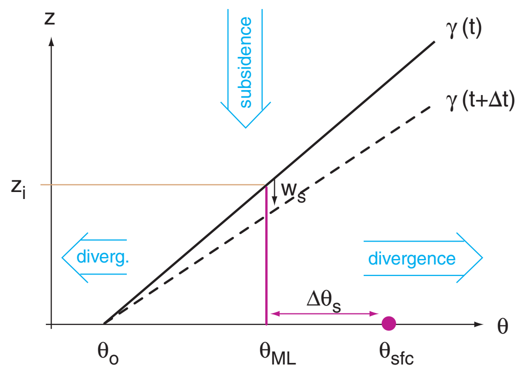
During World War I, Vilhelm Bjerknes, a Norwegian physicist with expertise in radio science and fluid mechanics, was asked in 1918 to form a Geophysical Institute in Bergen, Norway. Cut-off from weather data due to the war, he arranged for a dense network of 60 surface weather stations to be installed. Some of his students were C.-G. Rossby, H. Solberg, T. Bergeron, V. W. Ekman, H. U. Sverdrup, and his son Jacob Bjerknes.

Jacob Bjerknes used the weather station data to identify and classify cold, warm, and occluded fronts. He published his results in 1919, at age 22. The term "front" supposedly came by analogy to the battle fronts during the war. He and Solberg also later explained the life cycle of cyclones. Their description is known as the **Norwegian cyclone model**.

INFO • Warm Airmass Genesis

Modeling warm airmass creation (genesis) is an exercise in atmospheric boundary-layer (ABL) evolution. Since we do not cover ABLs in detail until a later chapter, the details are relegated to this INFO Box. You can safely skip them now, and come back later after you have studied ABLs.

Define a relative potential temperature θ based on a reference height (z) at the surface (z = 0). Namely, $\theta \approx T + \Gamma_d \cdot z$, where Γ_d = 9.8 °C km^{-1} is the dry adiabatic lapse rate.



Suppose that a cool, statically stable layer of air initially (subscript o) has a near-surface temperature of θ_o and a linear potential-temperature gradient $\gamma_o = \gamma = \Delta\theta/\Delta z$ before it comes to rest over a warm surface of temperature θ_{sfc} (see Fig. above). We want to predict the time evolution of the depth (z_i) and potential temperature (θ_{ML}) of this new airmass. Because this airmass is an ABL, we can use boundary-layer equations to predict z_i (the height of the convective mixed-layer ,ML), and θ_{ML} (the ML potential temperature).

ML depth increases by amount Δz_{ie} during a time interval Δt due to thermodynamic encroachment (i.e., warming under the capping sounding). This is an entrainment process that adds air to the ML through the ML top. Large scale **divergence** $\beta = \Delta U / \Delta x + \Delta V / \Delta y$ removes air horizontally from the ML and causes a subsidence velocity of magnitude w_s at the ML top:

$$w_s = \beta \cdot z_i^* \tag{7}$$

Thus, a change in ML depth results from the competition of these two terms:

$$z_i = z_i^* + \Delta z_{ie} - w_s \cdot \Delta t \tag{12}$$

where the asterisk * indicates a value from the previous time step.

The amount of heat ΔQ (as an incremental accumulated kinematic heat flux) transferred from the warm surface to the cooler air during time interval Δt under light-wind conditions depends on the temperature difference at the surface $\Delta\theta_s$ and the intensity of turbulence, as quantified by a buoyancy velocity scale w_b:

$$\Delta Q = b \cdot w_b \cdot \Delta\theta_s \cdot \Delta t \tag{10}$$

where b = 5x10^{-4} (dimensionless) is a convective heat transport coefficient (see the Heat chapter).

(continues in next column)

INFO *(continuation)*

The buoyancy velocity scale is:

$$w_b = \left[\frac{|g|}{\theta_{ML}^*} \cdot z_i^* \cdot \Delta\theta_s \right]^{1/2} \tag{9}$$

where $|g|$ = 9.8 m s^{-2} is gravitational acceleration.

That heat goes to warming θ_{ML}, which by geometry adds a trapezoidal area under the γ curve :

$$\Delta z_{ie} = \left[\frac{2 \cdot \Delta Q}{\gamma} + \left(z_i^*\right)^2 \right]^{1/2} - z_i^* \tag{11}$$

Knowing the entrainment rate, we get $\Delta\theta_{ML}$ geometrically from where it intercepts the γ curve:

$$\theta_{ML} = \theta_{ML}^* + \gamma \cdot \Delta z_{ie} \tag{13}$$

With this new ML temperature, we can update the surface temperature difference

$$\Delta\theta_s = \theta_{sfc} - \theta_{ML}^* \tag{8}$$

Knowing the large-scale divergence, we can also update the potential temperature profile in the air above the ML:

$$\gamma = \gamma_o \cdot \exp(\beta \cdot t) \tag{6}$$

All that remains is to update the time variable:

$$t = t^* + \Delta t \tag{5}$$

You might have noticed that some of the equations above are initially singular, when the ML has zero depth. So, for the first small time step ($\Delta t \approx$ 6 minutes = 360 s), you should use the following special equations in the following order:

$$t = \Delta t \tag{1}$$

$$\Delta\theta_s = \theta_{sfc} - \theta_o \tag{2}$$

$$z_i = \Delta\theta_s \cdot \left[\frac{2 \cdot b \cdot \Delta t}{\gamma_o} \cdot \left(\frac{|g|}{\theta_o} \right)^{1/2} \right]^{2/3} \tag{3}$$

$$\theta_{ML} = \theta_o + \gamma_o \cdot z_i \tag{4}$$

Then, for all the subsequent time steps, use the set of equations (5 to 13 in the order as numbered) to find the resulting ML evolution. Repeat eqs. (5 to 13) for each subsequent step. As the solution begins to change more gradually, you may use larger time steps Δt. The result is a toy model that describes warm airmass formation as an evolving convective boundary layer.

The Sample Application on the next page shows how this can be done with a computer spreadsheet. First, you need to specify the imposed constants γ_o , θ_o, β and θ_{sfc}. Next, initialize the values of: $\gamma = \gamma_o$, z_i = 0, and $\theta_{ML} = \theta_o$ at $t = 0$. Then, solve the equations for the first step. Finally, continue iterating for subsequent time steps to simulate warm airmass genesis.

These simulations show that greater divergence causes a shallower ML that can warm faster. Greater static stability in the ambient environment reduces the peak ML depth.

Sample Application (§)

Air of initial ML potential temperature 10°C comes to rest over a 20°C sea surface. Divergence is 10^{-6} s^{-1}, and the initial $\Delta\theta/\Delta z = \gamma = 3.3$ K km^{-1}. Find and plot the warm airmass evolution of potential temperature and depth.

Find the Answer

Given: θ_o = 10°C = 283K, θ_{sfc} = 20°C, γ_o = 3.3 K km^{-1}, $\beta = 10^{-6}$ s^{-1}.
Find: $\theta_{ML}(t)$ = ? °C, $z_i(t)$ = ? m

For the first time step of Δt = 6 min (=360 s), use eqs. (1 to 4 from the INFO Box on the previous page). For subsequent steps, repeatedly use eqs. (5 to 13 from that same INFO Box).

t (s)	t (d)	γ (K m^{-1})	w_s (m s^{-1})	$\Delta\theta_s$ (°C)	w_b (m s^{-1})	ΔQ (K·m)	Δz_{ie} (m)	z_i (m)	θ_{ML} (°C)
0	0.000							0	10
360	0.004			10				74	10.2
720	0.008	0.00330	0.00007	9.75	5.0	8.8	29.8	104	10.3
1080	0.013	0.00330	0.00010	9.66	5.9	10.3	26.4	131	10.4
1440	0.017	0.00330	0.00013	9.57	6.6	11.3	24.0	155	10.5
. . .									
259200	3.00	0.00428	0.00179	2.41	12.0	104.3	13.6	1789	17.7
270000	3.13	0.00432	0.00179	2.35	11.9	150.8	19.4	1789	17.7
280800	3.25	0.00437	0.00179	2.26	11.7	142.8	18.2	1788	17.8
. . .									
1684800	19.50	0.01779	0.00057	0.14	1.6	4.9	0.5	542	19.9
1728000	20.00	0.01858	0.00054	0.13	1.5	4.4	0.4	519	19.9

Sample results from the computer spreadsheet are shown above. The final answer is plotted in Fig. 12.5.

Check: Units OK. Physics OK. Fig. 12.5 reasonable.
Exposition: I used small time steps of Δt = 6 minutes initially, and then as Δz_{ie} became smaller, I gradually increased past Δt = 3 h to Δt = 12 h. The figure shows rapid initial modification of the cold, statically stable airmass toward a warm, unstable airmass. Maximum z_i is reached in about τ = 3.06 days (see INFO box), in agreement with Fig. 12.5.

INFO • Time of Max Airmass Thickness

The time τ to reach the peak ML thickness for warm airmass genesis is roughly

$$\tau \approx c \cdot \left(\frac{\theta_o}{|g| \cdot \beta \cdot \gamma_o \cdot e^{\beta \cdot \tau}} \right)^{1/3} \qquad \text{(c)}$$

where c = 140 (dimensionless), and the other variables are defined in the text. τ is typically about 1 to 4 days. Any further lingering of the airmass over the same surface temperature results in a loss of airmass thickness due to divergence.

Equation (c) is an implicit equation; namely, you need to know τ in order to solve for τ. Although this equation is difficult to solve analytically, you can iterate to quickly converge to a solution in about 5 steps in a computer spreadsheet. Namely, start with τ = 0 as the first guess and plug into the right side of eq. (c). Then solve for τ on the left side. For the next iteration, take this new τ and plug it in on the right, and solve for an updated τ on the left. Repeat until the value of τ converges to a solution; namely, when $\Delta\tau/\tau < \varepsilon$ for ε = 0.01 or smaller.

Sample Application (§)

For the conditions of the previous Sample Application, find the time τ that estimates when the new warm airmass has maximum thickness.

Find the Answer

Given: θ_o = 10°C = 283K, θ_{sfc} = 20°C,
γ_o = 3.3 K km^{-1}, $\beta = 10^{-6}$ s^{-1}.
Find: τ = ? days

Use eq. (c) from the INFO Box at left. Start with τ = 0. The first iteration is:

$$\tau = 140 \cdot \left[\frac{283\text{K}}{(9.8\text{m/s}^2) \cdot (10^{-6}\text{s}^{-1}) \cdot (0.0033\text{K/m}) \cdot e^0} \right]^{1/3}$$

= 288498 s = 3.339 days
Subsequent iterations give: τ =
3.033 -> 3.060 -> 3.057 -> 3.058 -> **3.058 days**

Check: Units OK. Physics OK. Agrees with Fig. 12.5.
Exposition: Convergence was quick. From Fig. 12.5, the actual time of this peak thickness was between 3 and 3.125 days, so eq. (c) does a reasonable job.

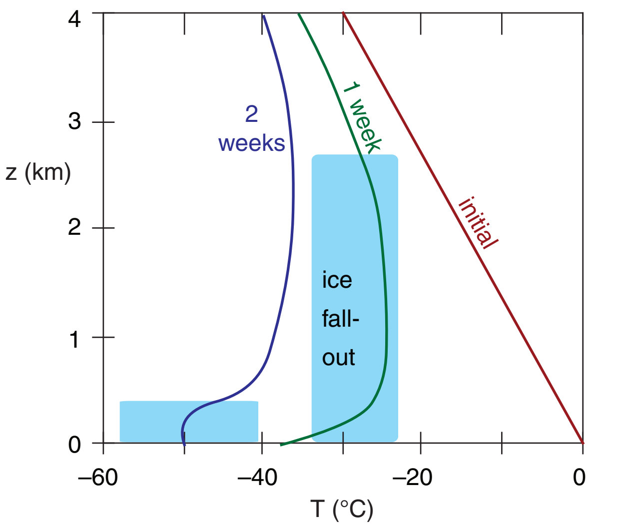


Figure 12.6
Genesis of a continental-polar air mass over arctic ice. The cloud/fog regions are shaded in light blue.

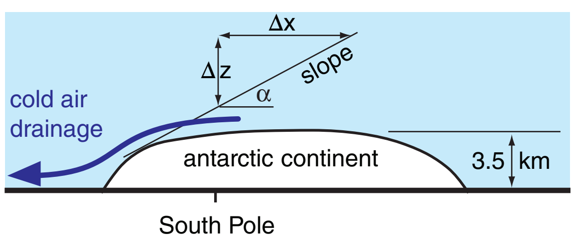


Figure 12.7
Cold katabatic winds (dark-blue arrow) draining from Antarctica.

Sample Application

Find the slope force per unit area acting on a katabatic wind of temperature –20°C with ambient air temperature 0°C. Assume a slope of $\Delta z/\Delta x = 0.1$.

Find the Answer

Given: $\Delta\theta = 20$ K, $T_e = 273$ K, $\Delta z/\Delta x = 0.1$
Find: $F_{x\ S}/m = ?$ m·s^{-2}

Use eq. (12.1):

$$\frac{F_{x\ S}}{m} = \frac{(9.8\text{m}\cdot\text{s}^{-2})\cdot(20\text{K})}{273\text{K}}\cdot(0.1)$$
$= \underline{\mathbf{0.072\ m{\cdot}s^{-2}}}$

Check: Units OK. Physics OK.
Exposition: This is two orders of magnitude greater than the typical synoptic forces (see the Dynamics chapter). Hence, drainage winds can be strong.

ice crystals to grow at the expense of evaporating liquid droplets, changing the cloud into an ice cloud.

Radiative cooling from cloud top creates cloudy "thermals" of cold air that sink, causing some turbulence that distributes the cooling over a deeper layer. Turbulent entrainment of air from above cloud top down into the cloud allows the cloud top to rise, and deepens the incipient airmass (Fig. 12.6).

The ice crystals within this cloud are so few and far between that the weather is described as cloudless ice-crystal precipitation. This can create some spectacular **halos** and other optical phenomena in sunlight (see the Atmospheric Optics chapter), including sparkling ice crystals known as **diamond dust**. Nevertheless, infrared radiative cooling in this cloudy air is much greater than in clear air, allowing the cooling rate to increase to 3°C day^{-1} over a layer as deep as 4 km.

During the two-week formation of this **continental-polar** or **continental-arctic** airmass, most of the ice crystals precipitate out leaving a thinner cloud of 1 km depth. Also, subsidence within the high pressure reduces the thickness of the cloudy airmass and causes some warming to partially counteract the radiative cooling.

Above the final fog layer is a nearly isothermal layer of air 3 to 4 km thick that has cooled about 30°C. Final air-mass temperatures are often in the range of –30 to –50 °C, with even colder temperatures near the surface.

While the Arctic surface consists of relatively flat sea-ice (except for Greenland), the Antarctic has mountains, high ice-fields, and significant surface topography (Fig. 12.7). As cold air forms by radiation, it can drain downslope as a **katabatic wind** (see the Regional Winds chapter). Steady winds of 10 m s^{-1} are common in the Antarctic interior, with speeds of 50 m s^{-1} along some of the steeper slopes.

As will be shown in the Regional Winds chapter, the buoyancy force per unit mass on a surface of slope $\Delta z/\Delta x$ translates into a quasi-horizontal slope-force per mass of:

$$\frac{F_{x\ S}}{m} = \frac{|g|\cdot\Delta\theta}{T_e}\cdot\frac{\Delta z}{\Delta x} \qquad \bullet(12.1)$$

$$\frac{F_{y\ S}}{m} = \frac{|g|\cdot\Delta\theta}{T_e}\cdot\frac{\Delta z}{\Delta y} \qquad \bullet(12.2)$$

where $|g|$ = 9.8 m·s^{-2} is gravitational acceleration, and $\Delta\theta$ is the potential-temperature difference between the draining cold air and the ambient air above. The ambient-air absolute temperature is T_e.

The sign of these forces should be such as to accelerate the wind downslope. The equations above work when the magnitude of slope $\Delta z/\Delta x$ is small, because then $\Delta z/\Delta x = \sin(\alpha)$, where α is the slope angle of the topography.

The katabatic wind speed in the Antarctic also depends on turbulent drag force against the ice surface, ambient pressure-gradient force associated with synoptic weather systems, Coriolis force, and turbulent drag caused by mixing of the draining air with the stationary air above it. At an average drainage velocity of 5 m s^{-1}, air would need over 2 days to move from the interior to the periphery of the continent, which is a time scale on the same order as the inverse of the Coriolis parameter. Hence, Coriolis force cannot be neglected.

Katabatic drainage removes cold air from the genesis regions and causes turbulent mixing of the cold air with warmer air aloft. The resulting cold-air mixture is rapidly distributed toward the outside edges of the antarctic continent.

One aspect of the global circulation is a wind that blows around the poles. This is called the **polar vortex**. Katabatic removal of air from over the antarctic reduces the troposphere depth, enhancing the persistence and strength of the antarctic polar vortex due to potential vorticity conservation.

12.2.2. Movement

Airmasses do not remain stationary over their birth place forever. After a week or two, a transient change in the weather pattern can push the airmass toward new locations.

When airmasses move, two things can happen:
(1) As the air moves over surfaces with different characteristics, the airmass begins to change. This is called **airmass modification**, and is described in the next subsection.
(2) An airmass can encounter another airmass. The boundary between these two airmasses is called a **front**, and is a location of strong gradients of temperature, humidity, and other airmass characteristics. Fronts are described in detail later.

Tall mountain ranges can strongly **block** or **channel** the movement of airmasses, because airmasses occupy the bottom of the troposphere. Fig. 12.8 shows a simplified geography of major mountain ranges.

For example, in the middle of North America, the lack of any major east-west mountain range allows the easy movement of cold polar air from Canada toward warm humid air from the Gulf of Mexico. This sets the stage for strong storms (see the Extratropical Cyclone chapter and the chapters on Thunderstorms).

The long north-south barrier of mountains (Rockies, Sierra Nevada, Cascades, Coast Range) along the west coast of North America impedes the easy entry of Pacific airmasses toward the center of

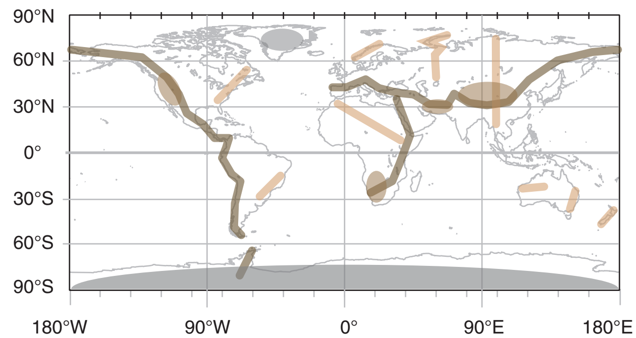


Figure 12.8
Locations of major mountain ranges (brown lines), lesser ranges (tan lines), and large plateaus of land or ice (tan ovals).

that continent. Those mountains also help protect the west coast from the temperature extremes experienced by the rest of the continent.

In Europe, the mountain orientation is the opposite. The Alps and the Pyrenees are east-west mountain ranges that inhibit movement of Mediterranean airmasses from reaching northward. The lack of major north-south ranges in west and central Europe allows the easy movement of maritime airmasses from the Atlantic to sweep eastward, bringing cool wet conditions.

One of the greatest ranges is the Himalaya Mountains, running east-west between India and China. Maritime tropical airmasses moving in from the Indian Ocean reach these mountains, causing heavy rains over India during the monsoon. The same mountains block the maritime air from reaching further northward, leaving a very dry Tibetan Plateau and Gobi Desert in its **rain shadow**.

The discussion above focused on blocking and channeling by the mountains. In some situations air can move over mountain tops (Fig. 12.9). When this happens, the airmass is strongly modified, as described next.

12.2.3. Modification

As an airmass moves from its origin, it is modified by the new landscapes under it. For example, a polar airmass will warm and gain moisture as it moves equatorward over warmer vegetated ground. Thus, it gradually loses its original identity.

12.2.3.1. Via Surface Fluxes

Heat and moisture transfer at the surface can be described with **bulk-transfer relationships** such as eq. (3.34). If we assume for simplicity that wind-induced turbulence creates a well-mixed airmass of quasi-constant thickness z_i, then the change

Sample Application

An mP airmass initially has T = 5°C & RH = 100%. Use a thermo diagram to find T & RH at: 1 Olympic Mtns. (elevation ≈ 1000 m), 2 Puget Sound (0 m), 3 Cascade Mtns (1500 m), 4 the Great Basin (500 m), 5 Rocky Mtns (2000 m), 6 the western Great Plains (1000 m).

Find the Answer

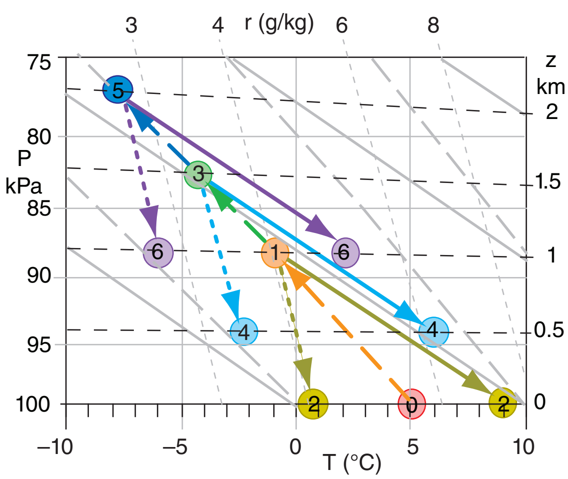
Given: Elevations from west to east (m) =
0, 1000, 0, 1500, 500, 2000, 1000
Initially RH=100%. Thus $T_d = T$ = 5°C
Find: T (°C) & RH (%) at surface locations 1 to 6.
Assume: All condensation precipitates out. No additional heat or moisture transfer from the surface. Start with air near sea level, where P_{SL} = 100 kPa.

Use Fig. 12.9 and an emagram from the Stability chapter. On the thermo diagram, the air parcel follows the following route: 0 - 1 - 2 - 1 - 3 - 4 - 3 - 5 - 6.

Initially (point 0), $T_d = T$ = 5°C. Because this air is already saturated, it would follow a saturated adiabat from point 0 to point 1 at z = 1 km, where the still-saturated air has $T_d = T$ = –1°C.

If all condensates precipitate out, then air would descend dry adiabatically (with T_d following an isohume) from point 1 to 2, giving T = 9°C and T_d = 1°C.

When this unsaturated air rises, it first does so dry adiabatically until it reaches its LCL at point 1. This now-cloudy air continues to rise toward point 3 moist adiabatically, where it has $T_d = T$ = –4°C. Etc.



Results, where Index = circled numbers in Fig above.

Index	z (km)	T (°C)	T_d (°C)	RH (%)
0	0	5	5	100
1	1	–1	–1	100
2	0	9	1	55
3	1.5	–4	–4	100
4	0.5	6	–2	54
5	2	–7	–7	100
6	1	2	–6	53

Check: Units OK. Physics OK. Figure OK.
Exposition: The airmass has lost its maritime identity by the time it reaches the Great Plains, and little moisture remains. Humidity for rain in the plains comes from the southeast (not from the Pacific).

of airmass potential temperature θ_{ML} with travel-distance Δx is:

$$\frac{\Delta\theta_{ML}}{\Delta x} \approx \frac{C_H \cdot (\theta_{sfc} - \theta_{ML})}{z_i} \qquad (12.3)$$

where C_H ≈ 0.01 is the bulk-transfer coefficient for heat (see the chapters on Heat and on the Atmospheric Boundary Layer).

If the surface temperature is horizontally homogeneous, then eq. (12.3) can be solved for the airmass temperature at any distance x from its origin:

$$\theta_{ML} = \theta_{sfc} - (\theta_{sfc} - \theta_{ML\,o}) \cdot \exp\left(-\frac{C_H \cdot x}{z_i} \right) \qquad (12.4)$$

where $\theta_{ML\,o}$ is the initial airmass potential temperature at location x = 0.

Sample Application

A polar airmass with initial θ = –20°C and depth = 500 m moves southward over a surface of 0°C. Find the initial rate of temperature change with distance.

Find the Answer

Given: θ_{ML} = –20°C, θ_{sfc} = 0°C, z_i = 500 m
Find: $\Delta\theta_{ML} / \Delta x$ = ? °C km^{-1}. Assume no mountains.

Use eq. (12.3):

$$\frac{\Delta\theta_{ML}}{\Delta x} \approx \frac{(0.01)\cdot[0°C - (-20°C)]}{(500m)} = \underline{\mathbf{0.4\ °C\ km^{-1}}}$$

Check: Units OK. Physics OK.
Exposition: Neither this answer nor eq. (12.4) depend on wind speed. While faster speeds give faster position change, they also cause greater heat transfer to/from the surface (eq. 3.34). These 2 effects cancel.

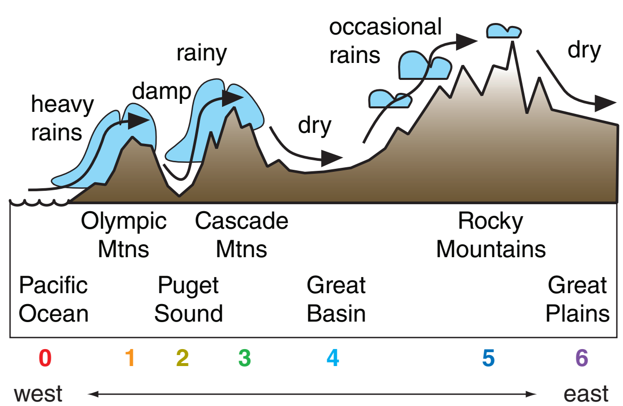


Figure 12.9
Modification of a Pacific airmass by flow over mountains in the northwestern USA. (Numbers are used in the Sample Application.)

12.2.3.2. Via Flow Over Mountains

If an airmass is forced to rise over mountain ranges, the resulting condensation, precipitation, and latent heating will dry and warm the air. For example, an airmass over the Pacific Ocean near the northwestern USA is often classified as maritime polar (mP), because it is relatively cool and humid. As the prevailing westerly winds move this airmass over the Olympic Mountains (a coastal mountain range), the Cascade Mountains, and the Rocky Mountains, there is substantial precipitation and latent heating (Fig. 12.9).

12.3. SURFACE FRONTS

Surface fronts mark the boundaries between airmasses at the Earth's surface. They usually have the following attributes:

- strong horizontal temperature gradient
- strong horizontal moisture gradient
- strong horizontal wind gradient
- strong vertical shear of the horizontal wind
- relative minimum of pressure
- high vorticity
- **confluence** (air converging horizontally)
- clouds and precipitation
- high static stability
- kinks in isopleths on weather maps

In spite of this long list of attributes, fronts are usually labeled by the surface temperature change associated with frontal passage.

Some weather features exhibit only a subset of attributes, and are not labeled as fronts. For example, a **trough** (pronounced like "trof") is a line of low pressure, high vorticity, clouds and possible precipitation, wind shift, and confluence. However, it often does not possess the strong horizontal temperature and moisture gradients characteristic of fronts.

Another example of an airmass boundary that is often not a complete front is the **dryline**. It is discussed later in this chapter.

Recall from the Weather Reports and Map Analysis chapter that fronts are always drawn on the warm side of the frontal zone. The frontal symbols (Fig. 12.10) are drawn on the side of the frontal line toward which the front is moving. For a stationary

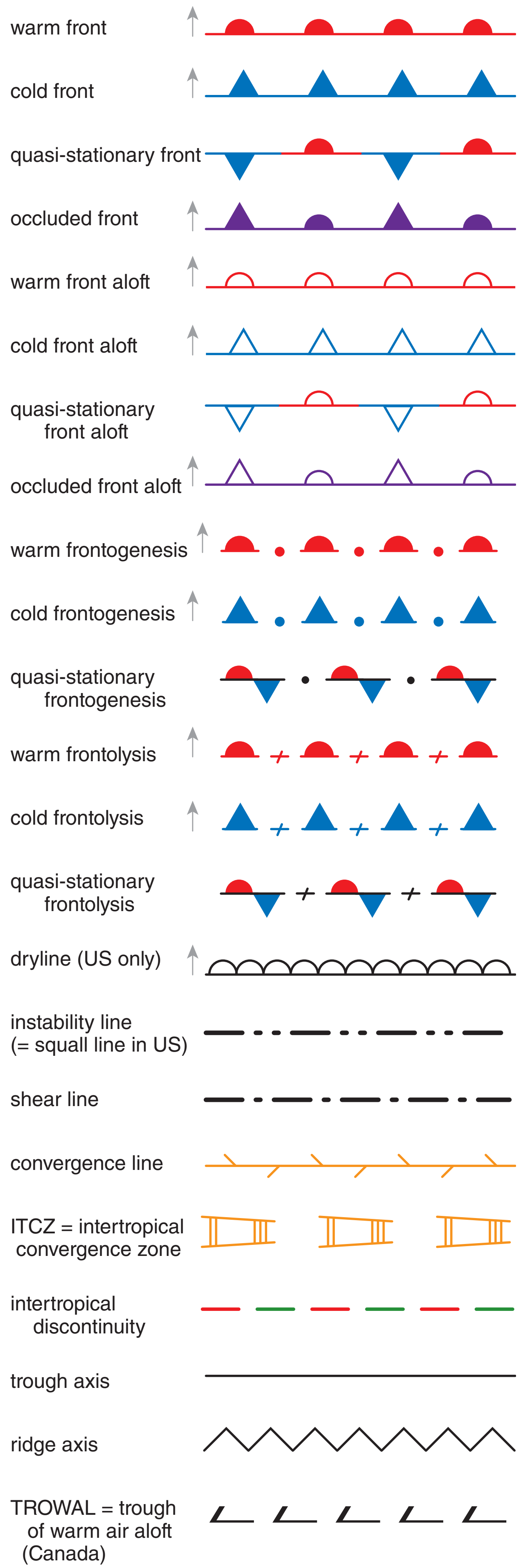
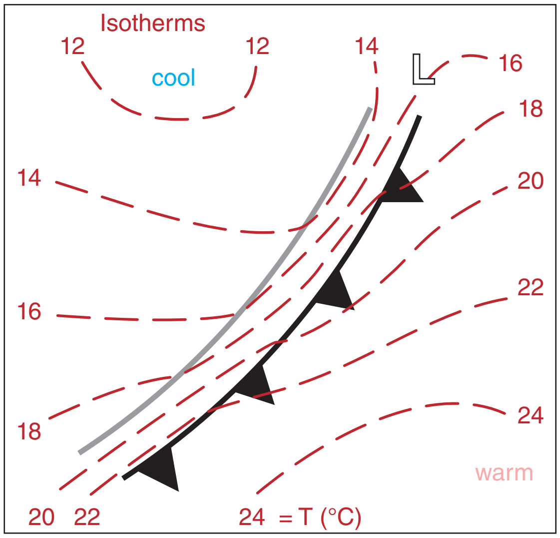
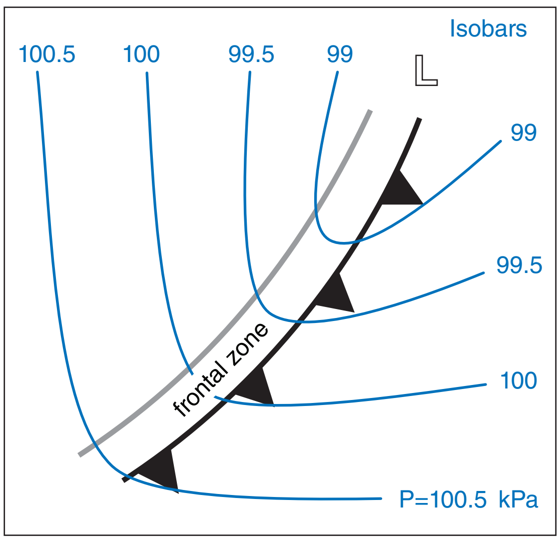


Figure 12.10 (at right)
Glyphs for fronts, other airmass boundaries, and axes (copied from the Weather Reports and Map Analysis chapter). The suffix "genesis" implies a forming or intensifying front, while "lysis" implies a weakening or dying front. A quasi-stationary front is a frontal boundary that doesn't move very much. Occluded fronts and drylines will be explained later.





front, the symbols on both sides of the frontal line indicate what type of front it would be if it were to start moving in the direction the symbols point.

Fronts are three dimensional. To help picture their structure, we next look at horizontal and vertical cross sections through fronts.

12.3.1. Horizontal Structure

12.3.1.1. Cold Fronts (Fig. 12.11)

In central N. America, winds ahead of cold fronts typically have a southerly component, and can form strong low-level jets at night and possibly during day. Warm, humid, hazy air advects from the south.

Sometimes a **squall line** of thunderstorms will form in advance of the front, in the warm air. These squall lines can be triggered by wind shear and by the kinematics (advection) near fronts. They can also consist of thunderstorms that were initially formed on the cold front, but progressed faster than the front.

Along the front are narrow bands of towering cumuliform clouds with possible thunderstorms and scattered showers. Along the front the winds are stronger and gusty, and pressure reaches a relative minimum. Thunderstorm anvils often spread hundreds of kilometers ahead of the surface front.

Winds shift to a northerly direction behind the front, advecting colder air from the north. This air is often clean with excellent visibilities and clear blue skies during daytime. If sufficient moisture is present, scattered cumulus or broken stratocumulus clouds can form within the cold airmass.

As this airmass consists of cold air advecting over warmer ground, it is statically unstable, con-

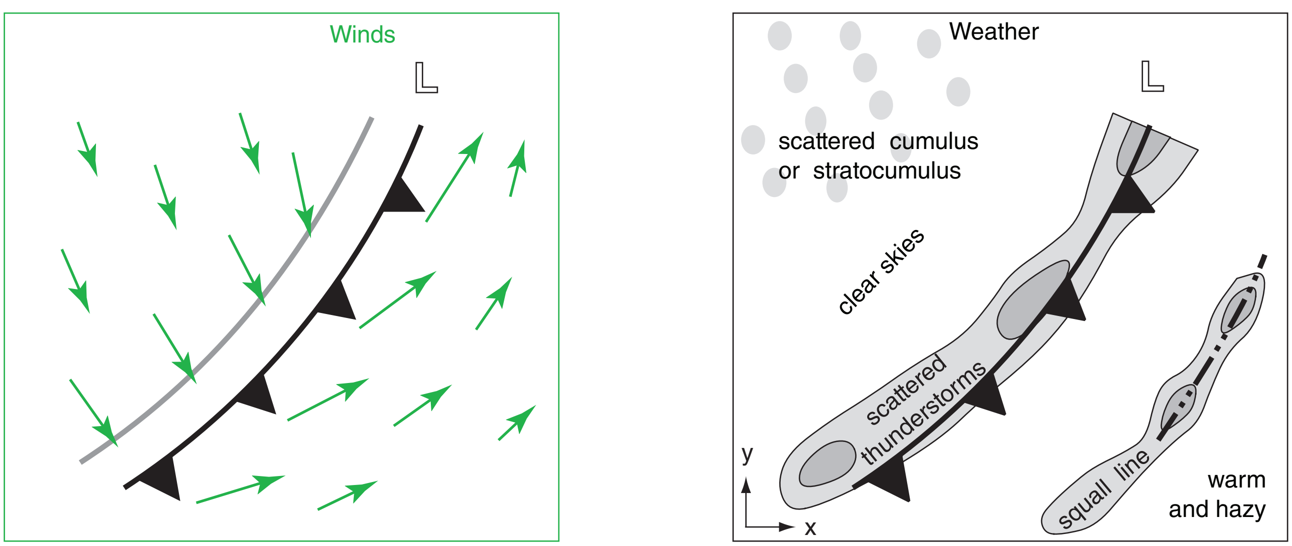


Figure 12.11 (both columns) *Cold front horizontal structure (maps can be overlain).*

vective, and very turbulent. However, at the top of the airmass is a very strong stable layer along the frontal inversion that acts like a lid to the convection. Sometimes over ocean surfaces the warm moist ocean leads to considerable post-frontal deep convection.

The idealized picture presented in Fig. 12.11 can differ considerably in the mountains.

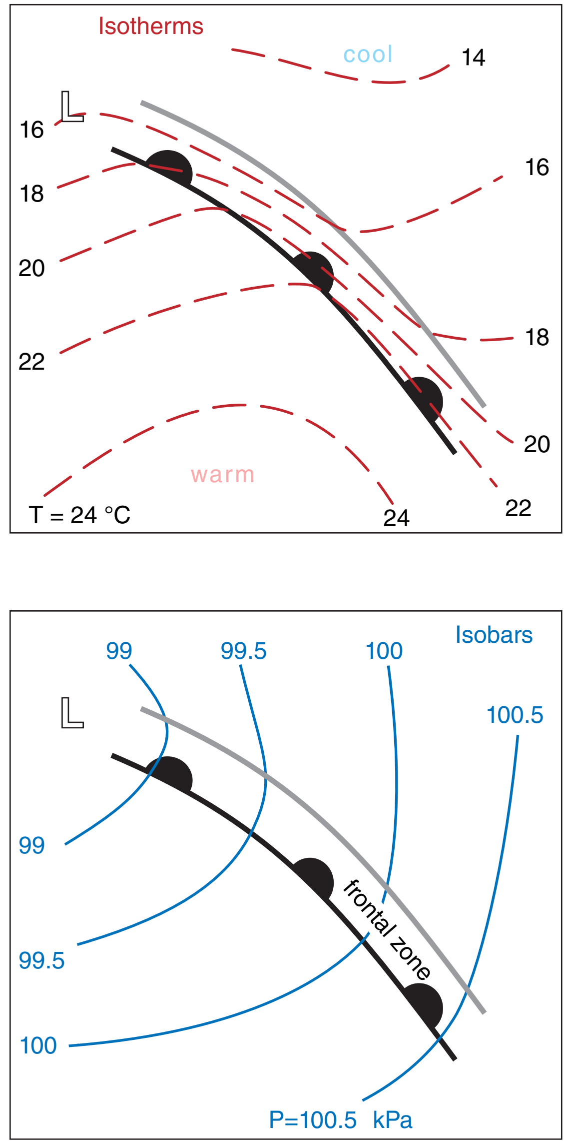
12.3.1.2. Warm Fronts (Fig. 12.12)

In central N. America, southeasterly winds ahead of the front bring in cool, humid air from the Atlantic Ocean, or bring in mild, humid air from the Gulf of Mexico.

An extensive deck of stratiform clouds (called a **cloud shield**) can occur hundreds of kilometers ahead of the surface front. In the cirrostratus clouds at the leading edge of this cloud shield, you can sometimes see halos, sundogs, and other optical phenomena. The cloud shield often wraps around the poleward side of the low center.

Along the frontal zone can be extensive areas of low clouds and fog, creating hazardous travel conditions. Nimbostratus clouds cause large areas of drizzle and light continuous rain. Moderate rain can form in multiple **rain bands** parallel to the front. The pressure reaches a relative minimum at the front.

Winds shift to a more southerly direction behind the warm front, advecting in warm, humid, hazy air. Although heating of air by the surface might not be strong, any clouds and convection that do form can often rise to relatively high altitudes because of weak static stabilities throughout the warm airmass.



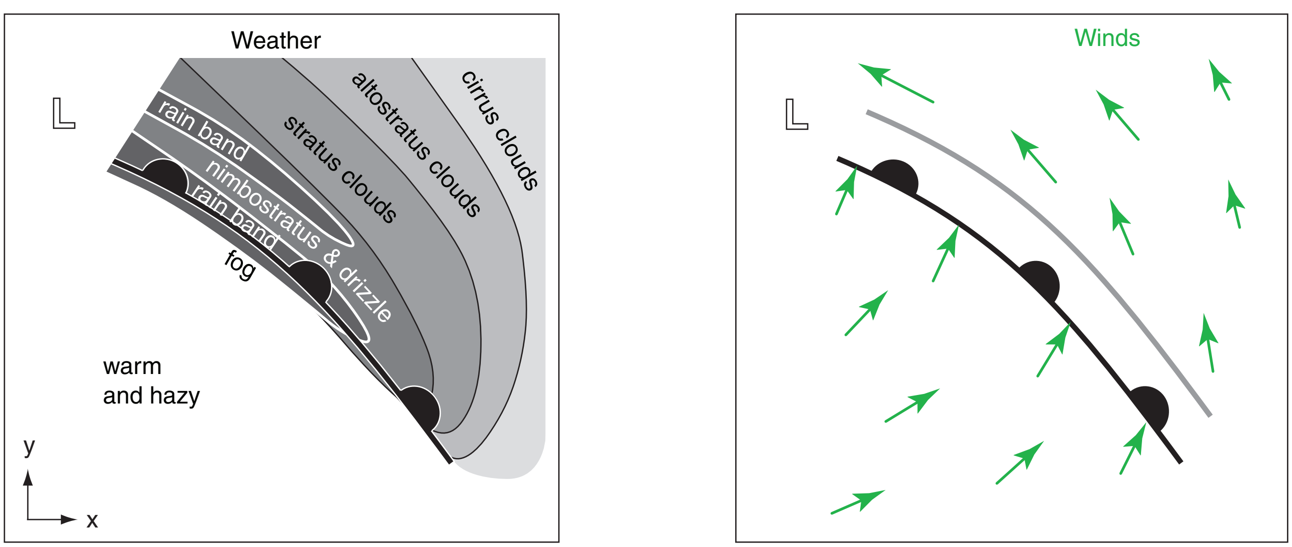
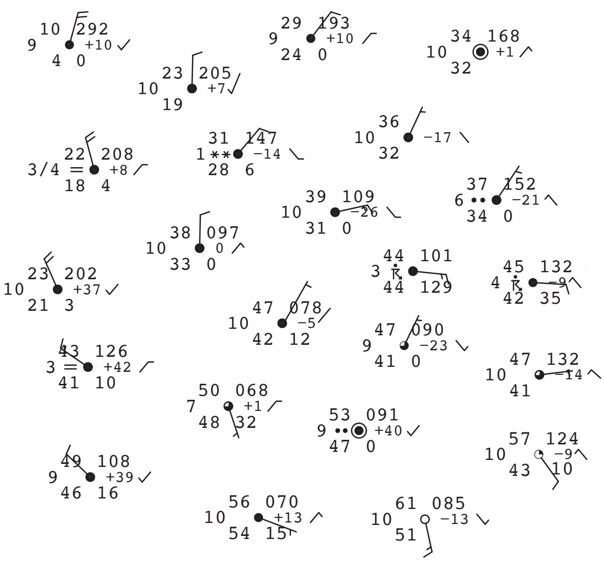


Figure 12.12 (both columns) *Warm front structure (maps can be overlain).*

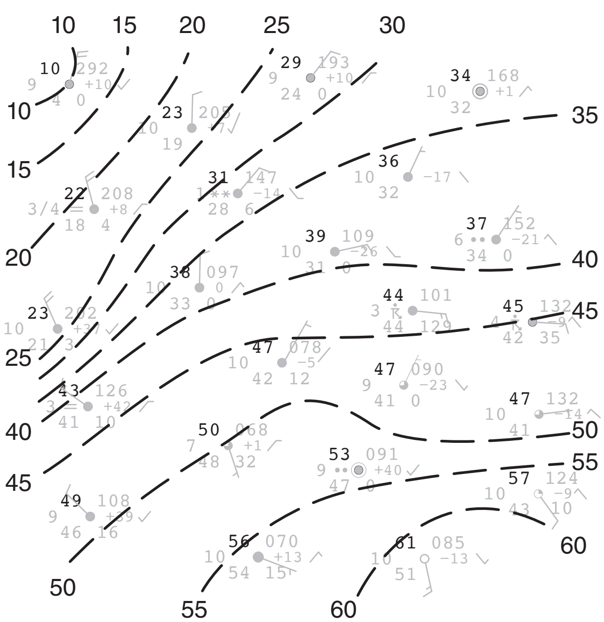
Sample Application

Given the plotted surface weather data below, analyze it for temperature (50°F and every 5°F above and below) and pressure (101.2 kPa and every 0.4 kPa above and below). Identify high- and low-pressure centers and fronts. Discuss how the winds, clouds and weather compare to the descriptions in Figs. 12.11 & 12.



Find the Answer

First, using methods shown in the Weather Reports & Map Analysis chapter, draw the isotherms (°F):

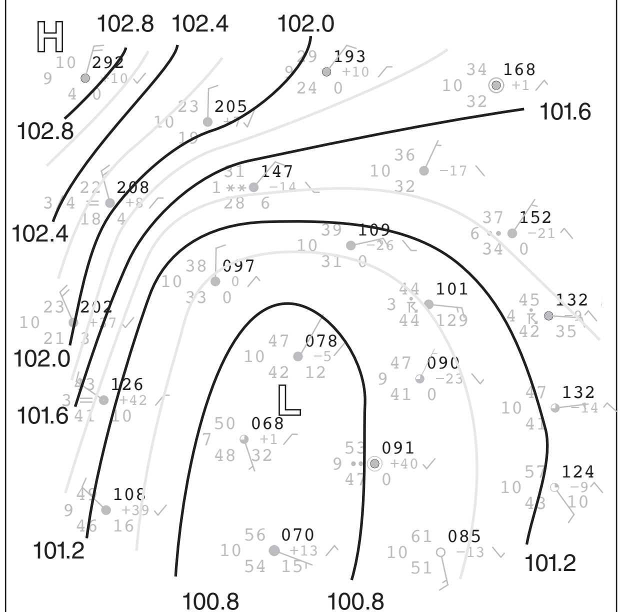


Next, analyze it for pressure. Recall that the plotted pressure is abbreviated. We need to prefix 9 or 10 to the left of the pressure code, and insert a decimal point two places from the right (to get kPa). Choose between 9 and 10 based on which one results in a pressure closest to standard sea-level pressure 101.3 kPa.

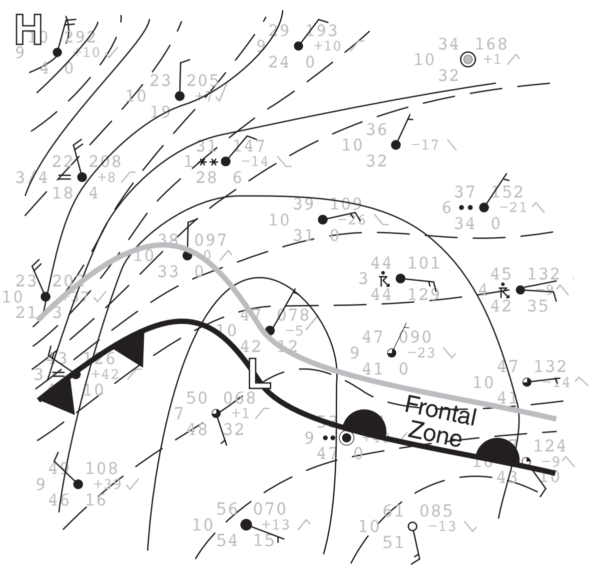
(continues in next column)

Sample Application *(continued)*

For the pressure data in this Sample Application, every prefix is 10. For example, the plotted pressure code 097 means 100.97 kPa. Similarly, 208 means 102.08 kPa. By analyzing pressures, we get the following map of isobars (kPa):



Next, overlay the isobars and isotherms, and find the frontal zones (drawn with the thick black and grey lines) and fronts (thick black line).



Check: Looks reasonable.

Exposition: Winds are generally circulating counterclockwise around the low center (except the very light winds, which can be sporadic). Overcast skies cover most of the region, with some thunderstorms and rain north of the warm front, and snow further northwest of the low where the air is colder.

12.3.2. Vertical Structure

Suppose that radiosonde observations (**RAOBs**) are used to probe the lower troposphere, providing temperature profiles such as those in Fig. 12.13a. To locate fronts by their vertical cross section, first convert the temperatures into potential temperatures θ (Fig. 12.13b). Then, draw lines of equal potential temperature (**isentropes**). Fig. 12.13c shows isentropes re-drawn at 5°C intervals. Often isentropes are labeled in Kelvin.

A frontal inversion is where the isentropes are packed closely together (shaded in Fig. 12.13c). This concept applies to both warm and cold fronts.

In the absence of diabatic processes such as latent heating, radiative heating, or turbulent mixing, air parcels follow isentropes when they move adiabatically. For example, consider the θ = 35°C parcel that is circled in Fig. 12.13b above weather station B. Suppose this parcel starts to move westward toward C.

If the parcel were to be either below or above the 35°C isentrope at its new location above point C, buoyant forces would tend to move it vertically to the 35°C isentrope. Such forces happen continuously while the parcel moves, constantly adjusting the altitude of the parcel so it rides on the isentrope.

The net movement is westward and upward along the 35°C isentrope. Air parcels that are forced to rise along isentropic surfaces can form clouds and precipitation, given sufficient moisture. Similarly, air blowing eastward would move downward along the sloping isentrope.

In three dimensions, you can picture **isentropic surfaces** separating warmer θ aloft from colder θ below. Analysis of the flow along these surfaces provides a clue to the weather associated with the front. Air parcels moving adiabatically must follow the "topography" of the isentropic surface. This is illustrated in the Extratropical Cyclones chapter.

At the Earth's surface, the boundary between cold and warm air is the **surface frontal zone**. This is the region where isentropes are packed relatively close together (Figs. 12.13b & c). The top of the cold air is called the **frontal inversion** (Fig. 12.13c). The frontal inversion is also evident at weather stations C and D in Fig. 12.13a, where the temperature increases with height. Frontal inversions of warm and cold fronts are gentle and of similar temperature change.

Within about 200 m of the surface, there are appreciable differences in frontal slope. The cold front has a steeper nose (slope ≈ 1 : 100) than the warm front (slope ≈ 1 : 300), although wide ranges of slopes have been observed.

Fronts are defined by their temperature structure, although many other quantities change across the front. Advancing cold air at the surface defines

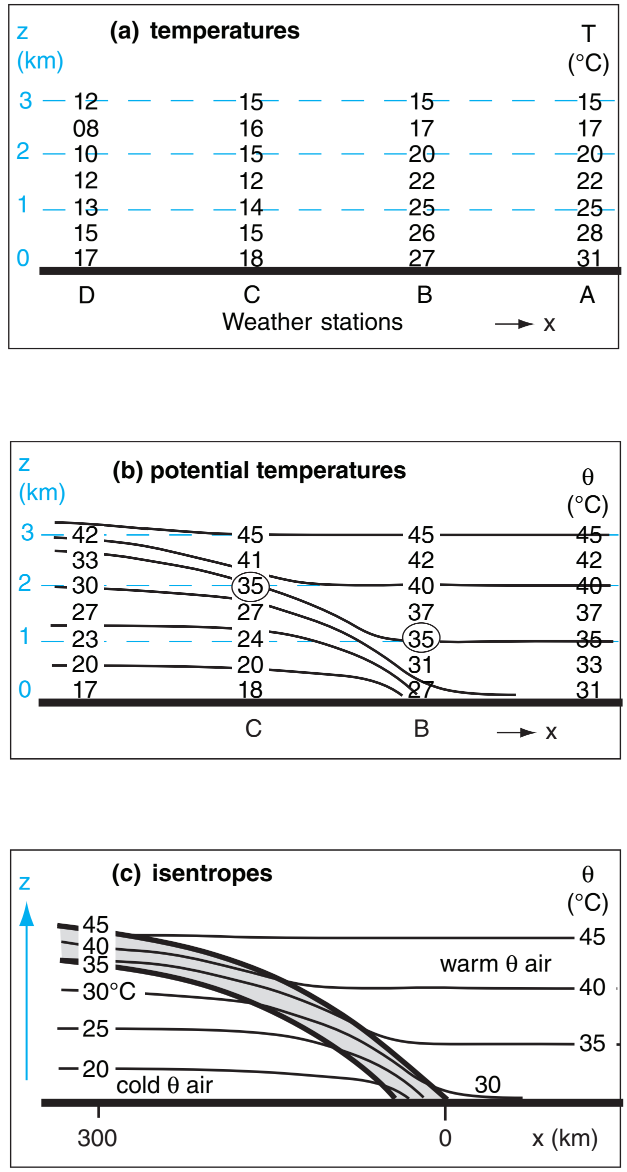


Figure 12.13
Analysis of soundings to locate fronts in a vertical cross section. Frontal zone / frontal inversion is shaded in bottom figure, and is located where the isentropes are tightly packed (close to each other).

Sample Application
What weather would you expect with a warm katafront? (See next page.)

Find the Answer & Exposition
Cumuliform clouds and showery precipitation would probably be similar to those in Fig. 12.16a, except that the bad weather would move in the direction of the warm air at the surface, which is the direction the surface front is moving.

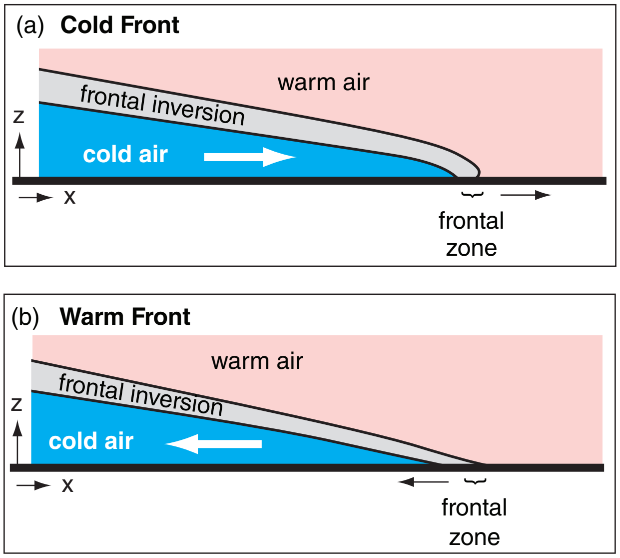


Figure 12.14
Vertical structure of fronts, based on cold air movement.

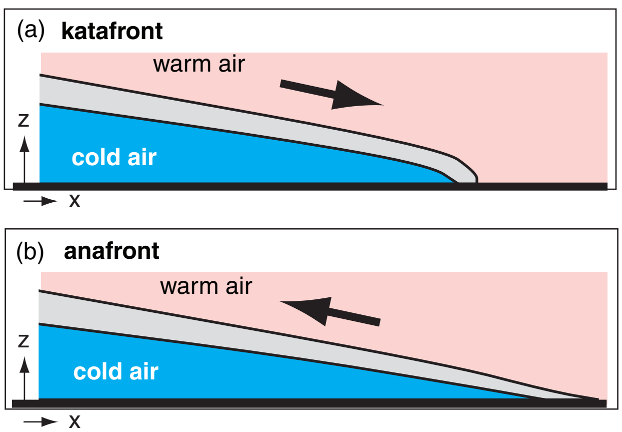


Figure 12.15
Vertical structure of fronts, based on overlying-air movement.

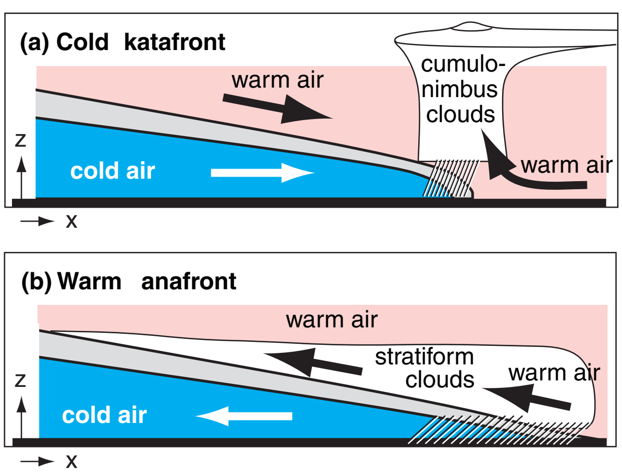


Figure 12.16
Typical vertical structure of fronts in central N. America.

the **cold front**, where the front moves toward the warm airmass (Fig. 12.14a). Retreating cold air defines the **warm front**, where the front moves toward the cold airmass (Fig. 12.14b).

Above the frontal inversion, if the warm air flows down the frontal surface, it is called a **katafront**, while warm air flowing up the frontal surface is an **anafront** (Fig. 12.15). It is possible to have cold katafronts, cold anafronts, warm katafronts, and warm anafronts.

Frequently in central N. America, the cold fronts are katafronts, as sketched in Fig. 12.16a. For this situation, warm air is converging on both sides of the frontal zone, forcing the narrow band of cumuliform clouds that is typical along the front. It is also common that warm fronts are anafronts, which leads to a wide region of stratiform clouds caused by the warm air advecting up the isentropic surfaces (Fig. 12.16b).

A **stationary front** is like an anafront where the cold air neither advances nor retreats.

12.4. GEOSTROPHIC ADJUSTMENT – PART 3

12.4.1. Winds in the Cold Air

Why does cold dense air from the poles not spread out over more of the Earth, like a puddle of water? Coriolis force is the culprit, as shown next.

Picture two air masses initially adjacent (Fig. 12.17a). The cold airmass has initial depth H and uniform virtual potential temperature θ_{v1}. The warm airmass has uniform virtual potential temperature $\theta_{v1} + \Delta\theta_v$. The average absolute virtual temperature is $\overline{T_v}$. In the absence of rotation of the coordinate system, you would expect the cold air to spread out completely under the warm air due to buoyancy, reaching a final state that is horizontally homogeneous.

However, on a rotating Earth, the cold air experiences Coriolis force (to the right in the Northern Hemisphere) as it begins to move southward. Instead of flowing across the whole surface, the cold air spills only distance a before the winds have turned 90°, at which point further spreading stops (Fig. 12.17b).

At this quasi-equilibrium, pressure-gradient force associated with the sloping cold-air interface balances Coriolis force, and there is a steady geostrophic wind U_g from east to west. The process of approaching this equilibrium is called **geostrophic adjustment**, as was discussed in the previous chapter. Real atmospheres never quite reach this equilibrium.

At equilibrium, the final spillage distance a of the front from its starting location equals the **external Rossby-radius of deformation**, λ_R:

$$a = \lambda_R = \frac{\sqrt{|g| \cdot H \cdot \Delta\theta_v / \overline{T_v}}}{f_c} \qquad \bullet(12.5)$$

where f_c is the Coriolis parameter and $|g|$ = 9.8 m s^{-2} is gravitational acceleration magnitude.

The geostrophic wind U_g in the cold air at the surface is greatest at the front (neglecting friction), and exponentially decreases behind the front:

$$U_g = -\sqrt{|g| \cdot H \cdot (\Delta\theta_v / \overline{T_v})} \cdot \exp\left(-\frac{y+a}{a}\right) \qquad \bullet(12.6)$$

for $-a \le y \le \infty$. The depth of the cold air h is:

$$h = H \cdot \left[1 - \exp\left(-\frac{y+a}{a}\right)\right] \qquad \bullet(12.7)$$

which smoothly increases to depth H well behind the front (at large y).

Figs. 12.17 are highly idealized, having airmasses of distinctly different temperatures with a sharp interface in between. For a fluid with a smooth continuous temperature gradient, geostrophic adjustment occurs in a similar fashion, with a final equilibrium state as sketched in Fig. 12.18. The top of this diagram represents the top of the troposphere, and the top wind vector represents the jet stream.

This state has high surface pressure under the cold air, and low surface pressure under the warm air (see the General Circulation chapter). On the cold side, isobaric surfaces are more-closely spaced in height than on the warm side, due to the hypsometric relationship. This results in a pressure reversal aloft, with low pressure (or low heights) above the cold air and high pressure (or high heights) above the warm air.

Horizontal pressure gradients at low and high altitudes create opposite geostrophic winds, as indicated in Fig. 12.18. Due to Coriolis force, the air represented in Fig. 12.18 is in equilibrium; namely, the cold air does not spread any further.

This behavior of the cold airmass is extremely significant. It means that the planetary-scale flow, which is in approximate geostrophic balance, is unable to complete the job of redistributing the cold air from the poles and warm air from the tropics. Yet some other process must be acting to complete the job of redistributing heat to satisfy the global energy budget (in the General Circulation chapter).

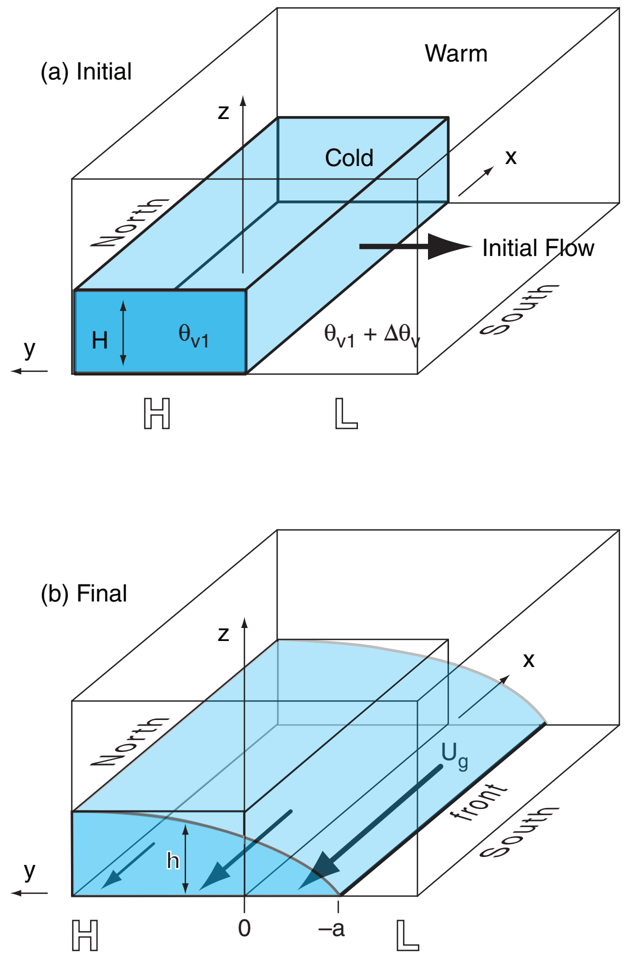


Figure 12.17
Geostrophic adjustment of a cold front. (a) Initial state. (b) Final state is in dynamic equilibrium, which is never quite attained in the real atmosphere. Cold air is shaded light blue.

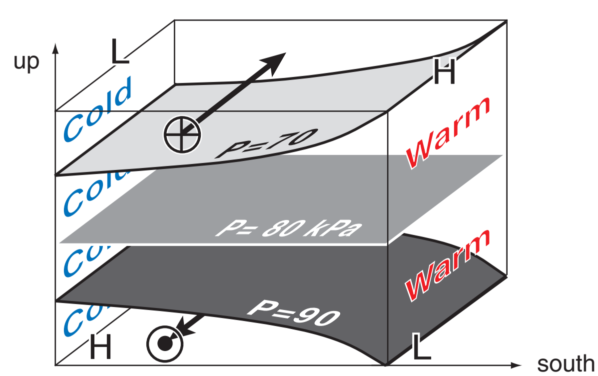


Figure 12.18
Final dynamic state after geostrophic adjustment within an environment containing continuous temperature gradients. Arrows represent geostrophic wind. Shaded areas are isobaric surfaces. H and L indicate high and low pressures relative to surrounding pressures at the same altitude. Dot-circle represents the tip of an arrow pointing toward the reader, x-circle represents the tail feathers of an arrow pointing into the page.

HIGHER MATH • Geostrophic Adjustment

We can verify that the near-surface (frictionless) geostrophic wind is consistent with the sloping depth of cold air. The geostrophic wind is related to the horizontal pressure gradient at the surface by:

$$U_g = -\frac{1}{\rho \cdot f_c} \cdot \frac{\Delta P}{\Delta y} \qquad (10.26a)$$

Assume that the pressure at the top of the cold airmass in Figs. 12.17 equals the pressure at the same altitude h in the warm airmass. Thus, surface pressures will be different due to only the difference in weight of air below that height.

Going from the top of the sloping cold airmass to the bottom, the vertical increase of pressure is given by the hydrostatic eq: $\Delta P_{cold} = -\rho_{cold} \cdot |g| \cdot h$

A similar equation can be written for the warm air below h. Thus, at the surface, the difference in pressures under the cold and warm air masses is:

$$\Delta P = -|g| \cdot h \cdot (\rho_{cold} - \rho_{warm})$$

multiplying the RHS by $\bar{\rho} / \bar{\rho}$, where $\bar{\rho}$ is an average density, yields:

$$\Delta P = -|g| \cdot h \cdot \bar{\rho} \cdot [(\rho_{cold} - \rho_{warm}) / \bar{\rho}]$$

As was shown in the Buoyancy section of the Stability chapter, use the ideal gas law to convert from density to virtual temperature, remembering to change the sign because the warmer air is less dense. Also, $\Delta T_v = \Delta\theta_v$. Thus: $\Delta P = -|g| \cdot h \cdot \bar{\rho} \cdot [(\theta_{v\ warm} - \theta_{v\ cold}) / \bar{T}_v]$ where this is the pressure change in the negative y direction.

Plugging this into eq. (10.26a) gives:

$$U_g = -\frac{|g| \cdot (\Delta\theta_v / T_v)}{f_c} \cdot \frac{\Delta h}{\Delta y}$$

which we can write in differential form:

$$\boxed{U_g = -\frac{|g| \cdot (\Delta\theta_v / T_v)}{f_c} \cdot \frac{\partial h}{\partial y}} \qquad (b)$$

The equilibrium value of h was given by eq. (12.7):

$$h = H \cdot \left[1 - \exp\left(-\frac{y+a}{a}\right)\right] \qquad (12.7)$$

Thus, the derivative is:

$$\frac{\partial h}{\partial y} = \frac{H}{a} \cdot \exp\left(-\frac{y+a}{a}\right)$$

Plugging this into eq. (b) gives:

$$U_g = -\frac{|g| \cdot (\Delta\theta_v / T_v)}{f_c} \cdot \frac{H}{a} \cdot \exp\left(-\frac{y+a}{a}\right) \qquad (c)$$

But from eq. (12.5) we see that:

$$f_c \cdot a = f_c \cdot \lambda_R = \sqrt{|g| \cdot H \cdot (\Delta\theta_v / T_v)} \qquad (12.5)$$

Eq. (c) then becomes:

$$\boxed{U_g = -\sqrt{|g| \cdot H \cdot (\Delta\theta_v / T_v)} \cdot \exp\left(-\frac{y+a}{a}\right)} \qquad (12.6)$$

Thus, the wind is consistent (i.e., geostrophically balanced) with the sloping height.

That other process is the action of cyclones and Rossby waves. Many small-scale, short-lived cyclones are not in geostrophic balance, and they can act to move the cold air further south, and the warm air further north. These cyclones feed off the potential energy remaining in the large-scale flow, namely, the energy associated with horizontal temperature gradients. Such gradients have potential energy that can be released when the colder air slides under the warmer air (see Fig. 11.1).

Sample Application

A cold airmass of depth 1 km and virtual potential temperature 0°C is embedded in warm air of virtual potential temperature 20°C. Find the Rossby deformation radius, the maximum geostrophic wind speed, and the equilibrium depth of the cold airmass at $y = 0$. Assume $f_c = 10^{-4}\ s^{-1}$.

Find the Answer

Given: $\overline{T_v} = 280$ K, $\Delta\theta_v = 20$ K, $H = 1$ km, $f_c = 10^{-4}\ s^{-1}$.

Find: a = ? km, U_g = ? m s^{-1} at $y = -a$, and h = ? km at $y = 0$.

Use eq. (12.5):

$$a = \frac{\sqrt{(9.8\text{m}\cdot\text{s}^{-2})\cdot(1000\text{m})\cdot(20\text{K})/(280\text{K})}}{10^{-4}\text{s}^{-1}}$$

= **265 km**

Use eq. (12.6):

$$U_g = -\sqrt{(9.8\text{m}\cdot\text{s}^{-2})(1000\text{m})(20\text{K})/(280\text{K})} \cdot \exp(0)$$

= **–26.5 m s^{-1}**

Use eq. (12.7):

$$h = (1\text{km}) \cdot [1 - \exp(-1)] = \mathbf{0.63\ km}$$

Check: Units OK. Physics OK.

Exposition: Frontal-zone widths on the order of a = 200 km are small compared to lengths (1000s km).

12.4.2. Winds in the Warm Over-riding Air

Across the frontal zone is a stronger-than-background horizontal temperature gradient. In many fronts, the horizontal temperature gradient is strongest near the surface, and weakens with increasing altitude.

The thermal-wind relationship tells us that the geostrophic wind will increase with height in strong horizontal temperature gradients. If the frontal zone extends vertically over a large portion of the troposphere, then the wind speed will continue to increase with height, reaching a maximum near the tropopause.

Thus, jet streams are associated with frontal zones. The jet blows parallel to the frontal zone, with greatest wind speeds on the warm side of the frontal zone. If the cold air is advancing as a cold front, then this jet is known as a **pre-frontal jet**.

This is illustrated in Fig. 12.19. Plotted are isentropes, isobars, isotachs, and the frontal zone. The cross-frontal direction is north-south in this figure, causing a pre-frontal jet from the West (blowing into the page, in this diagram).

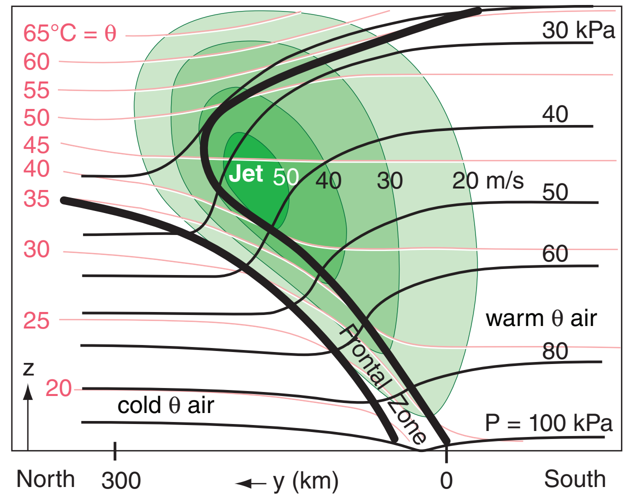


Figure 12.19
Vertical section across a cold front in the N. Hemisphere. Thick lines outline the frontal zone in the troposphere, and show the tropopause in the top of the graph. Medium black lines are isobars. Thin pink lines are isentropes. Green shaded areas indicate isotachs, for a jet that blows from the West (into the page).

12.4.3. Frontal Vorticity

Combining the cold-air-side winds from Fig. 12.17b and warm-air-side winds from Fig. 12.19 into a single diagram yields Fig. 12.20. In this sketch, the warm air aloft and south of the front has geostrophic winds from the west, while the cold air near the ground has geostrophic winds from the east.

Thus, across the front, $\Delta U_g/\Delta y$ is negative, which means that the relative vorticity (eq. 11.20) of the geostrophic wind is positive (ζ_r = +) at the front (grey curved arrows in Fig. 12.20). In fact, cyclonic vorticity is found along fronts of any orientation. Also, stronger density contrasts across fronts cause greater positive vorticity.

Also, frontogenesis (strengthening of a front) is often associated with horizontal convergence of air from opposite sides of the front (see next section). Horizontal convergence implies vertical divergence (i.e., vertical stretching of air and updrafts) along the front, as required by mass continuity. But stretching increases vorticity (see the chapters on Atmos. Stability, General Circulation, and Extratropical Cyclones). Thus, frontogenesis is associated with updrafts (and associated clouds and bad weather) and with increasing relative vorticity.

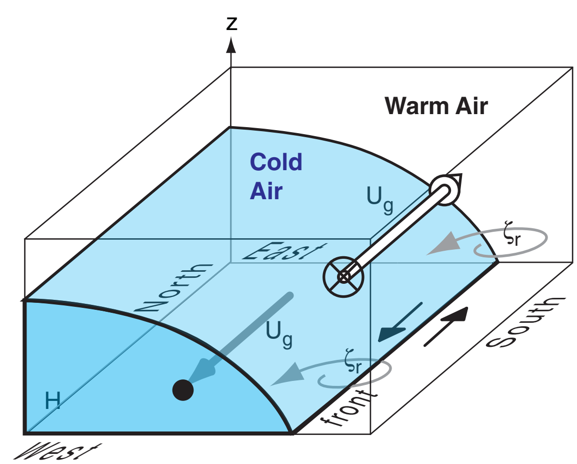


Figure 12.20
Sketch of cold front, combining winds (thick straight grey arrow) in the cold air (light blue shading) from Fig. 12.17b with the winds in the warm air (thick white arrow) from Fig. 12.19. These winds cause shear across the front, shown with the black arrows. This shear is associated with positive (cyclonic) relative vorticity of the geostrophic wind (thin curved grey arrows).

INFO • The Polar Front

Because of Coriolis force, cold arctic air cannot spread far from the poles, causing a quasi-permanent frontal boundary in the winter hemisphere. This is called the **polar front**.

It has a wavy irregular shape where some segments advance as cold fronts, other segments retreat as warm fronts, some are stationary, and others are weak and cause gaps in the front.

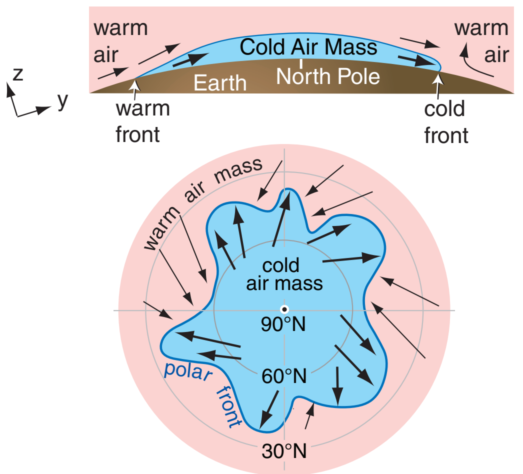


Figure a. *Top: Vertical slice through the North Pole. Bottom: View from space looking down on N. Pole.*

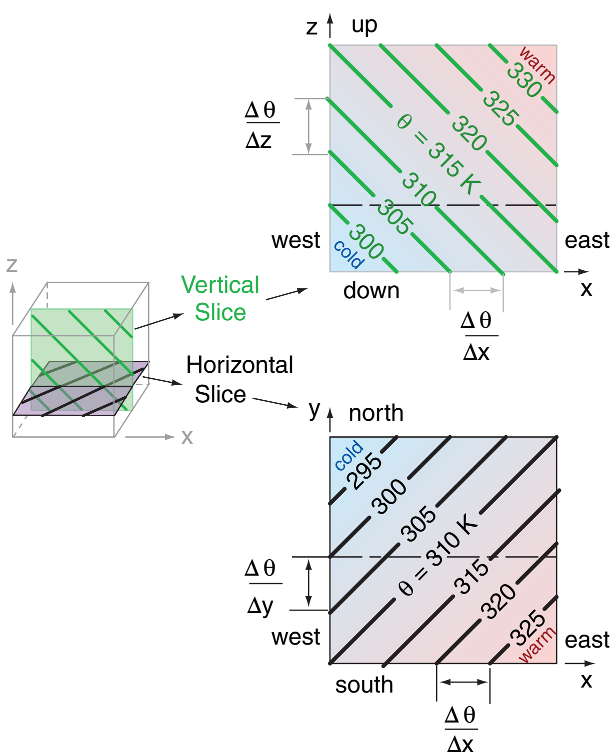


Figure 12.21
Vertical and horizontal slices through a volume of atmosphere, showing initial conditions prior to frontogenesis. The thin horizontal dashed lines show where the planes intersect.

12.5. FRONTOGENESIS

Fronts are recognized by the rapid change in surface temperature across the frontal zone. Hence, the horizontal temperature gradient (temperature change per distance across the front) is one measure of frontal strength. Usually potential temperature is used instead of temperature to simplify the problem when vertical motions can occur.

Physical processes that tend to increase the potential-temperature gradient are called **frontogenetic** — literally they cause the birth or strengthening of the front. Frontogenesis can be caused by kinematic, thermodynamic, and dynamic processes.

12.5.1. Kinematics

Kinematics refers to motion or advection, with no regard for driving forces. This class of processes cannot create potential-temperature gradients, but it can strengthen or weaken existing gradients.

From earlier chapters, we saw that radiative heating causes north-south temperature gradients between the equator and poles. Also, the general circulation causes the jet stream to meander, which creates transient east-west temperature gradients along troughs and ridges. The standard-atmosphere also has a vertical gradient of potential temperature in the troposphere (θ increases with z). Thus, it is fair to assume that temperature gradients often exist, which could be strengthened during kinematic frontogenesis.

To illustrate kinematic frontogenesis, consider an initial potential-temperature field with uniform gradients in the x, y, and z directions, as sketched in Fig. 12.21. The gradients have the following signs (for this particular example):

$$\frac{\Delta\theta}{\Delta x} = + \qquad \frac{\Delta\theta}{\Delta y} = - \qquad \frac{\Delta\theta}{\Delta z} = + \tag{12.8}$$

Namely, potential temperature increases toward the east, decreases toward the north, and increases upward. There are no fronts in this picture initially.

We will examine the subset of advections that tends to create a cold front aligned north-south. Define the strength of the front as the potential-temperature gradient across the front:

$$\text{Frontal Strength} = FS = \frac{\Delta\theta}{\Delta x} \qquad \bullet(12.9)$$

The change of frontal strength with time due to advection is given by the kinematic frontogenesis equation:

$$\frac{\Delta(FS)}{\Delta t} = -\left(\frac{\Delta\theta}{\Delta x}\right)\cdot\left(\frac{\Delta U}{\Delta x}\right) - \left(\frac{\Delta\theta}{\Delta y}\right)\cdot\left(\frac{\Delta V}{\Delta x}\right) - \left(\frac{\Delta\theta}{\Delta z}\right)\cdot\left(\frac{\Delta W}{\Delta x}\right) \quad \bullet(12.10)$$

Strengthening Confluence Shear Tilting

12.5.1.1. Confluence

Suppose there is a strong west wind U approaching from the west, but a weaker west wind departing at the east (Fig. 12.22 top). Namely, the air from the west almost catches up to air in the east.

For this situation, $\Delta U/\Delta x$ is negative , and $\Delta\theta/\Delta x$ is positive in eq. (12.10). Hence, the product of these two terms, when multiplied by the negative sign attached to the confluence term, tends to strengthen the front [$\Delta(FS)/\Delta t$ is positive]. In the shaded region of Fig. 12.22, the isentropes are packed closer together; namely, it has become a frontal zone.

12.5.1.2. Shear

Suppose the wind from the south is stronger on the east side of the domain than the west (Fig. 12.23 top). This is one type of wind shear. As the isentropes on the east advect northward faster than those on the west, the potential-temperature gradient is strengthened in-between, creating a frontal zone.

While the shear $\Delta V/\Delta x$ is positive, the northward temperature gradient $\Delta\theta/\Delta y$ is negative. Thus, the product is positive when the preceding negative sign from the shear term is included. Frontal strengthening occurs for this case [$\Delta(FS)/\Delta t$ is positive].

12.5.1.3. Tilting

If updrafts are stronger on the cold side of the domain than the warm side, then the vertical potential-temperature <u>gradient</u> will be tilted into the horizontal. The result is a strengthened frontal zone (Fig. 12.24).

The horizontal gradient of updraft velocity is negative in this example ($\Delta W/\Delta x$ = –), while the vertical potential-temperature gradient is positive ($\Delta\theta/\Delta z$ = +). The product, when multiplied by the negative sign attached to the tilting term, yields a positive contribution to the strengthening of the front for this case [$\Delta(FS)/\Delta t$ = +].

While this example was contrived to illustrate frontal strengthening, <u>for most real fronts, the tilting term causes weakening</u>. Such **frontolysis** is weakest near the surface because vertical motions are smaller there (the wind cannot blow through the ground).

Tilting is important and sometimes dominant for upper-level fronts, as described later.

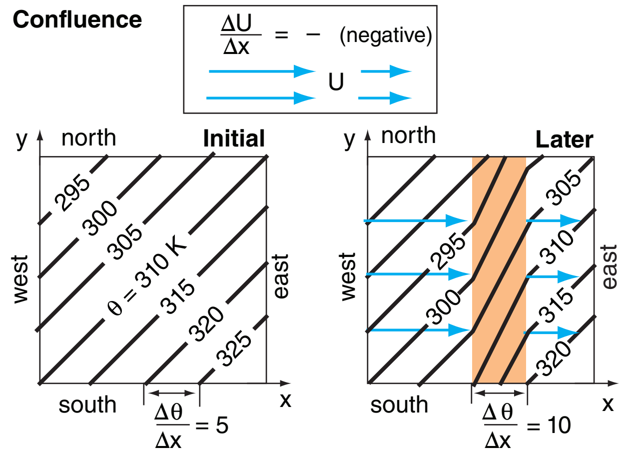


Figure 12.22
Confluence strengthens the frontal zone (shaded orange) in this case. Arrow tails indicate starting locations for the isotherms.

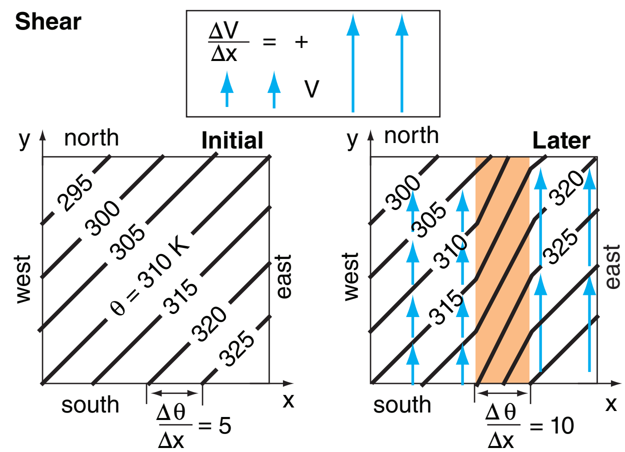


Figure 12.23
Shear strengthens the frontal zone (shaded orange) in this case. Arrow tails indicate starting locations for the isotherms.

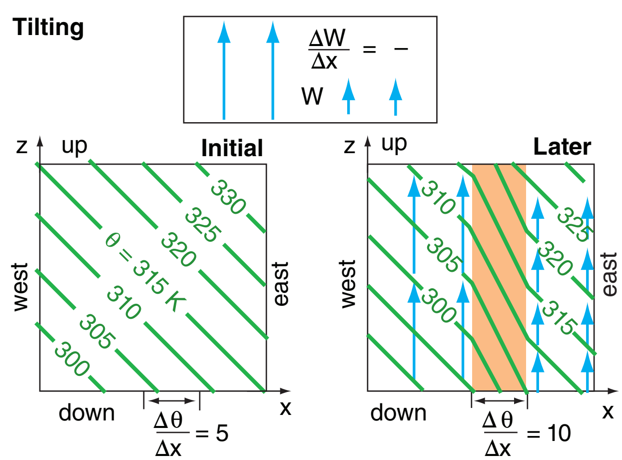


Figure 12.24
Tilting of the vertical temperature gradient into the horizontal strengthens the frontal zone (shaded orange) in this illustration.

Sample Application

Given an initial environment with $\Delta\theta/\Delta x$ = 0.01°C km^{-1}, $\Delta\theta/\Delta y$ = –0.01°C km^{-1}, and $\Delta\theta/\Delta z$ = 3.3°C km^{-1}. Also, suppose that $\Delta U/\Delta x$ = – 0.05 (m/s) km^{-1}, $\Delta V/\Delta x$ = 0.05 (m/s) km^{-1}, and $\Delta W/\Delta x$ = 0.02 (cm/s) km^{-1}. Find the kinematic frontogenesis rate.

Find the Answer

Given: (see above)

Find: $\Delta(FS)/\Delta t$ = ? °C·km^{-1}·day^{-1}

Use eq. (12.10):

$$\frac{\Delta(FS)}{\Delta t} = -\left(0.01\frac{°C}{km}\right)\cdot\left(-0.05\frac{m/s}{km}\right) - \left(-0.01\frac{°C}{km}\right)\cdot \left(0.05\frac{m/s}{km}\right) - \left(3.3\frac{°C}{km}\right)\cdot\left(0.0002\frac{m/s}{km}\right)$$

$$= +0.0005 + 0.0005 - 0.00066 \text{ °C·m·s}^{-1}\text{·km}^{-2}$$

$$= \underline{\mathbf{+0.029\ °C·km^{-1}·day^{-1}}}$$

Check. Units OK. Physics OK.

Exposition: Frontal strength $\Delta\theta/\Delta x$ nearly tripled in one day, increasing from 0.01 to 0.029 °C km^{-1}.

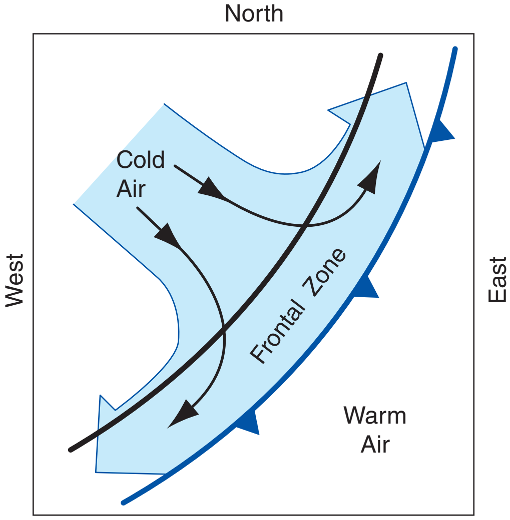


Figure 12.25

An illustration of the near-surface horizontal air-flow pattern at a frontal zone.

INFO • Confluence and Diffluence

Consider two neighboring streams of air, both moving in nearly the same dominant wind direction. Confluence is when those two streams become closer together. Diffluence is when the streams move further apart. Confluence/diffluence is analogous to streams of cars merging as they enter/exit an expressway or autobahn. In confluent flow, air is converging in a direction perpendicular to the dominant flow direction. Diffluence and divergence are similarly related.

12.5.1.4. Deformation

The previous figures presented idealized kinematic scenarios. Often in real fronts the flow field is a more complex combination of scenarios. For example, Fig. 12.25 shows a **deformation** (change of shape) flow field in the cold air, with **confluence** (→ ← coming together horizontally) of air perpendicular to the front, and **diffluence** (← → horizontal spreading of air) parallel to the front.

In such a flow field both convergence and shear affect the temperature gradient. For example, consider the two identical deformation fields in Figs. 12.26a & b, where the only difference is the angle of the isentropes in a frontal zone relative to the **axis of dilation** (the line toward which confluence points, and along which diffluence spreads).

For initial angles less than 45° (Fig. 12.26a & a'), the isentropes are pushed closer together (frontogenesis) and tilted toward a shallower angle. For initial angles greater than 45°, the isentropes are spread farther apart (frontolysis) and tilted toward a shallower angle. Using this info to analyze Fig. 12.25 where the isentropes are more-or-less parallel to the frontal zone (i.e., initial angle << 45°), we would expect that flow field to cause frontogenesis.

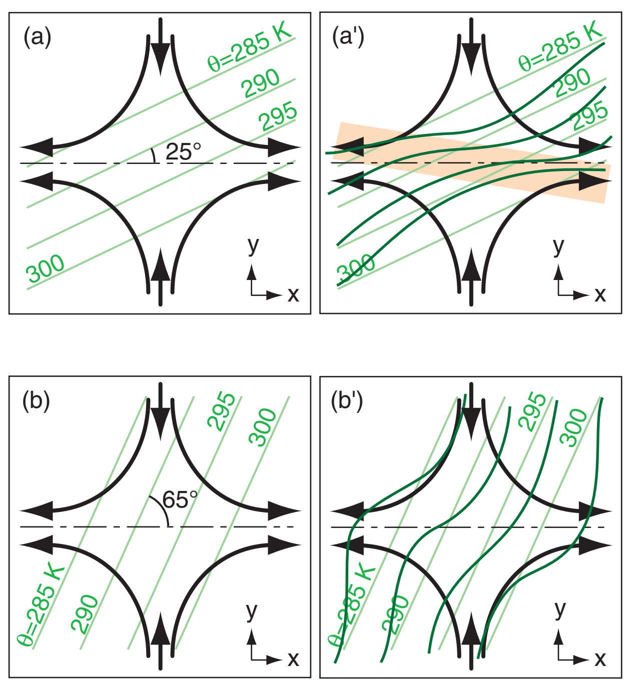


Figure 12.26

Black arrows are wind, and green lines are isentropes. Dashed black line is the axis of dilation. (a) and (a') are initial and final states for shallow initial angle, showing frontogenesis. The orange shaded area in (a') highlights the strengthened frontal zone. (b) and (b') are initial and final states for steep initial angle, showing frontolysis. (after J. Martin, 2006: "Mid-Latitude Atmospheric Dynamics: A First Course. Wiley.)

12.5.2. Thermodynamics

The previous kinematic examples showed **adiabatic** advection (potential temperature was conserved while being blown with the wind). However, **diabatic** (non-adiabatic) thermodynamic processes can heat or cool the air at different rates on either side of the domain. These processes include radiative heating/cooling, conduction from the surface, turbulent mixing across the front, and latent heat release/absorption associated with phase changes of water in clouds.

Define the diabatic warming rate (DW) as:

$$\text{Diabatic Warming Rate} = DW = \frac{\Delta\theta}{\Delta t} \qquad \bullet(12.11)$$

If diabatic heating is greater on the warm side of the front than the cold side, then the front will be strengthened:

$$\frac{\Delta(FS)}{\Delta t} = \frac{\Delta(DW)}{\Delta x} \qquad \bullet(12.12)$$

In most real fronts, **turbulent mixing** between the warm and cold sides weakens the front (i.e., causes **frontolysis**). **Conduction** from the surface also contributes to frontolysis. For example, behind a cold front, the cold air blows over a usually-warmer surface, which heats the cold air (i.e., airmass modification) and reduces the temperature contrast across the front. Similarly, behind warm fronts, the warm air is usually advecting over cooler surfaces.

Over both warm and cold fronts, the warm air is often forced to rise. This rising air can cause **condensation** and cloud formation, which strengthen fronts by warming the already-warm air.

Radiative cooling from the tops of stratus clouds reduces the temperature on the warm side of the front, contributing to frontolysis of warm fronts. Radiative cooling from the tops of post-cold frontal stratocumulus clouds can strengthen the front by cooling the already-cold air.

12.5.3. Dynamics

Kinematics and thermodynamics are insufficient to explain observed frontogenesis. While kinematic frontogenesis gives doubling or tripling of frontal strength in a day (see previous Sample Applications), observations show that frontal strength can increase by a factor of 15 during a day. Dynamics can cause this rapid strengthening.

Because fronts are long and narrow, we expect along-front flow to tend toward geostrophy, while across front flows could be ageostrophic. We can anticipate this by using the **Rossby number** (see

Sample Application

A thunderstorm on the warm side of a 200 km wide front rains at 2 mm h^{-1}. Find the frontogenesis rate.

Find the Answer

Given: $RR = 0$ at $x = 0$,
and $RR = 2$ mm h^{-1} at $x = 200$ km
Find: $\Delta(FS)/\Delta t$ = ? °C·km^{-1}·day^{-1}

From the Heat chapter:
$\Delta\theta/\Delta t = (0.33°\text{C mm}^{-1})\cdot RR = 0.66°\text{C h}^{-1}$.

Combine eqs. (12.11) & (12.12): $\Delta(FS)/\Delta t = \Delta(\Delta\theta/\Delta t)/\Delta x$

$\Delta(FS)/\Delta t = [(0.66°\text{C h}^{-1})\cdot(24 \text{ h day}^{-1}) - 0] \,/\, [200 \text{ km} - 0]$
$= (15.84°\text{C day}^{-1})/(200\text{km}) =$ **0.079 °C·km^{-1}·day^{-1}**

Check: Units OK. Physics OK. Magnitude good.
Exposition: This positive value indicates thermodynamic frontogenesis.

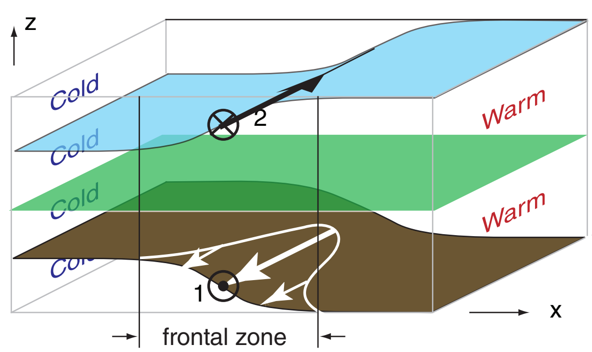


Figure 12.27
Isobaric surfaces near a hypothetical front, before being altered by dynamics. Vectors show equilibrium geostrophic winds, initially (state "o"). Dot-circle (1) represents the tip of an arrow pointing toward the reader; x-circle (2) represents the tail feathers of an arrow pointing into the page.

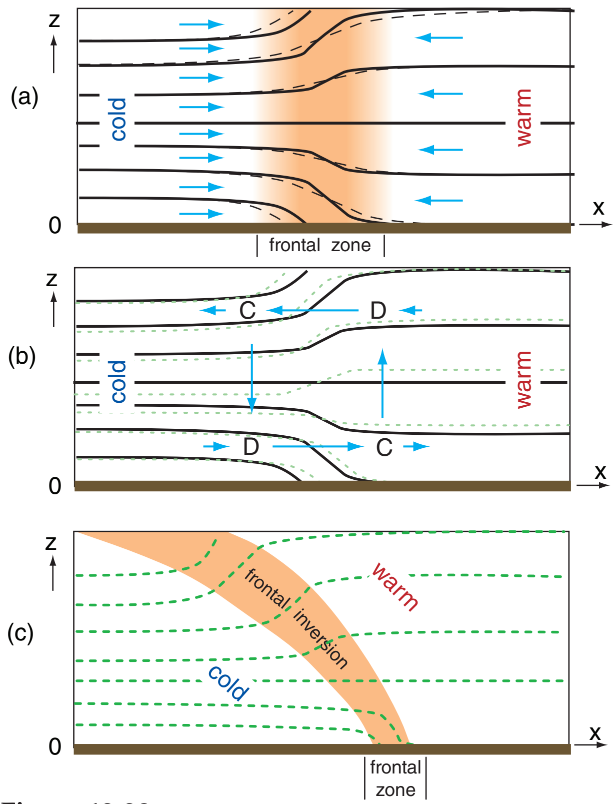


Figure 12.28
Vertical cross section showing dynamic strengthening of a front. The frontal zone is shaded orange; lines are isobars; and arrows are ageostrophic winds. D and C are regions of horizontal divergence and convergence. (a) Initial state (thin dashed lines) modified (thick black lines) by confluence (arrows). (b) Ageostrophic circulation, called a Sawyer-Eliassen circulation starts moving the black lines toward the dashed green lines. (c) Final equilibrium state.

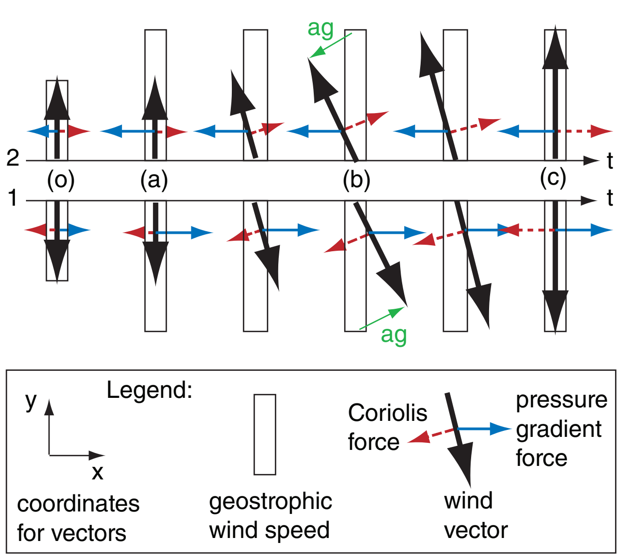


Figure 12.29
Time lines labeled 1 & 2 are for the vectors 1 & 2 in Fig. 12.27. Initial state (o) is given by Fig. 12.27. Later states (a)-(c) correspond to Figs. 12.28(a)-(c). The ageostrophic wind (ag) at time (b) is also indicated. States (o) and (c) are balanced.

INFO Box in the Forces & Winds chapter). Fronts can be of order 1000 km long, but of order 100 km wide. Thus, the Rossby number for along-front flow is of order $Ro = 0.1$. But the across-front Rossby number is of order $Ro = 1$. Recall that flows tend toward geostrophy when $Ro < 1$. Thus, ageostrophic dynamics are anticipated across the front, as are illustrated next.

Picture an initial state in geostrophic equilibrium with winds parallel to the front, as sketched in Fig. 12.27. This figure shows a special situation where pressure gradients and geostrophic winds exist only midway between the left and right sides. Zero gradients and winds are at the left and right sides. A frontal zone is in the center of this diagram.

Suppose some external forcing such as kinematic confluence due to a passing Rossby wave causes the front to strengthen a small amount, as sketched in Fig. 12.28a. Not only does the potential-temperature gradient tighten, but the pressure gradient also increases due to the hypsometric relationship.

The increased pressure gradient implies a different, increased geostrophic wind. However, initially, the actual winds are slower due to inertia, with magnitude equal to the original geostrophic speed.

While the actual winds adjust toward the new geostrophic value, they temporarily turn away from the geostrophic direction (Fig. 12.29) due to the imbalance between pressure-gradient and Coriolis forces. During this transient state (b), there is a component of wind in the x-direction (Fig. 12.28b). This is called **ageostrophic flow**, because there is no geostrophic wind in the x-direction.

Because mass is conserved, horizontal convergence and divergence of the U-component of wind cause vertical circulations. These are thermally direct circulations, with cold air sinking and warm air rising and moving over the colder air. The result is a temporary cross-frontal, or **transverse circulation** called a **Sawyer-Eliassen circulation**. The updraft portion of the circulation can drive convection, and cause precipitation.

The winds finally reach their new equilibrium value equal to the geostrophic wind. In this final state, there are no ageostrophic winds, and no cross-frontal circulation. However, during the preceding transient stage, the ageostrophic cross-frontal circulation caused extra dynamic confluence near the surface, which adds to the original kinematic confluence to strengthen the surface front. The transverse circulation also tilts the front (Fig. 12.28c).

In summary, a large and relatively steady geostrophic wind blows parallel to the front (Fig. 12.27). A weak, transient, cross-frontal circulation can be superimposed (Fig. 12.28b). These two factors are also important for upper-tropospheric fronts, as described later.

12.6. OCCLUDED FRONTS AND MID-TROPOSPHERIC FRONTS

When three or more airmasses come together, such as in an **occluded front**, it is possible for one or more fronts to ride over the top of a colder airmass. This creates lower- or mid-tropospheric fronts that do not touch the surface, and which would not be signaled by temperature changes and wind shifts at the surface. However, such **fronts aloft** can trigger clouds and precipitation observed at the surface.

Occluded fronts occur when cold fronts catch up to warm fronts. What happens depends on the temperature and static-stability difference between the cold advancing air behind the cold front and the cold retreating air ahead of the warm front.

Fig. 12.30 shows a **cold front occlusion**, where very cold air that is very statically stable catches up to, and under-rides, cooler air that is less statically stable. The warm air that was initially between these two cold airmasses is forced aloft. Most occlusions in interior N. America are of this type, due to the very cold air that advances from Canada in winter.

Observers at the surface would notice stratiform clouds in advance of the front, which would normally signal an approaching warm front. However, instead of a surface warm front, a surface occluded front passes, and the surface temperature decreases like a cold front. The trailing edge of cool air aloft marks the warm front aloft.

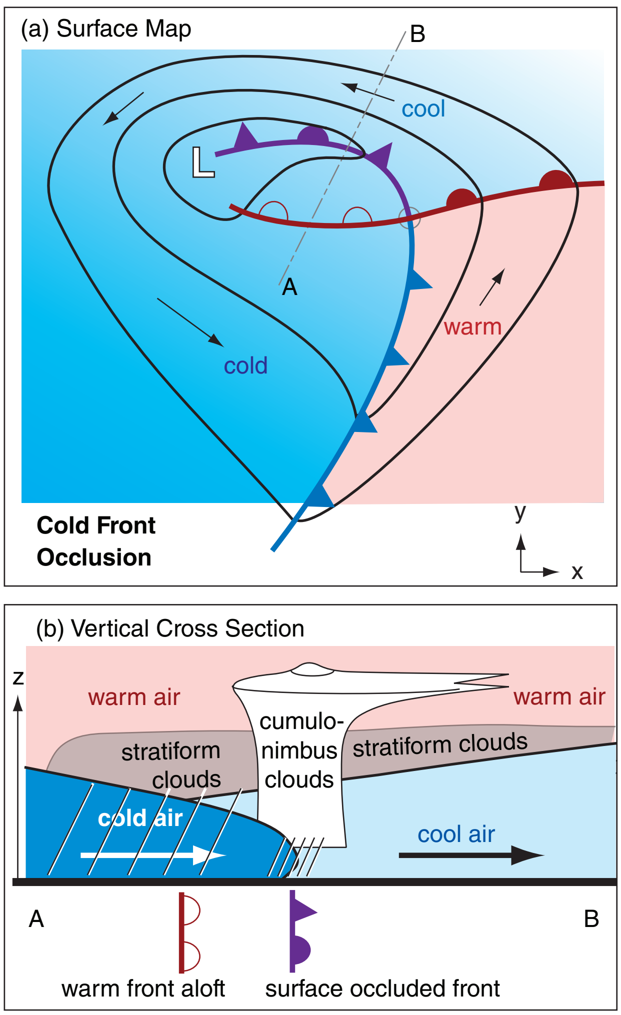


Figure 12.30
Cold front occlusion. (a) Surface map showing position of surface cold front (blue triangles), surface warm front (red semicircles), surface occluded front (purple triangles and semicircles), and warm front aloft (white semicircles). (b) Vertical cross section along slice A-B from top diagram. Diagonal lines = rain.

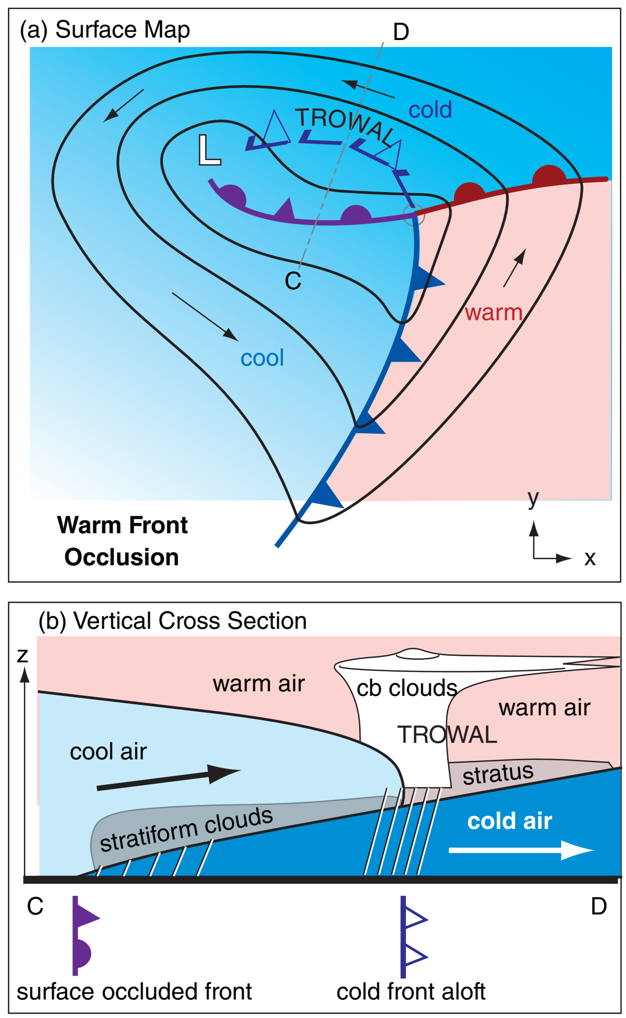


Figure 12.31
Warm front occlusion. (a) Surface map. Symbols are similar to Fig. 12.30, except that white triangles denote a cold front aloft. (b) Vertical cross section along slice C-D from top diagram. TROWAL = trough of warm air aloft. The open grey circle shows the triple-point location (meeting of 3 surface fronts).

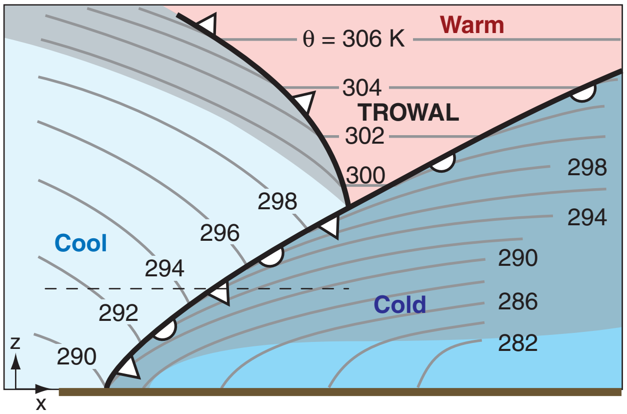


Figure 12.32
Idealized vertical cross-section through a warm front occlusion, showing the bottom 5 km of the troposphere. Grey lines are isentropes. Grey shading indicates strong static stability (regions where the isentropes are packed closer together in the vertical). TROWAL = trough of warm air aloft.

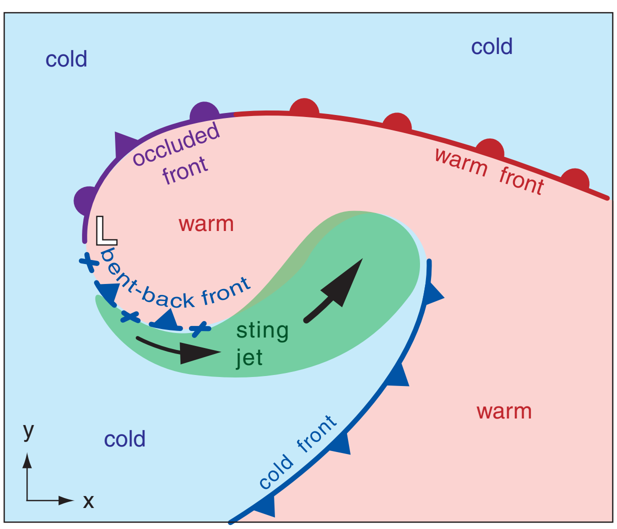


Figure 12.33
Idealized Shapiro-Keyser cyclone model for the N. Hemisphere, showing the warm, cold, occluded, and bent-back fronts. 𝕃 marks the low-pressure center (i.e., the core of the extratropical cyclone). Associated with some bent-back fronts is a sting jet of fast wind at the Earth's surface, where the air in this jet has descended from mid-tropospheric altitudes.

The location in Figs. 12.30 and 12.31 where the cold, warm, and occluded fronts intersect at the surface is called the **triple point**.

Fig. 12.31 shows a **warm front occlusion**, where cool air that is less statically stable catches up to, and over-rides, colder air that is more statically stable, forcing aloft the warm air that was in-between. Most occlusions in Europe and the Pacific Northwest USA are this type, due to mild cool air that advances from over the cool ocean during winter.

Observers at the surface notice stratiform clouds in advance of the front. But before the surface front arrives, there can be showers or thunderstorms associated with the cool front aloft. Later, a surface occluded front passes with widespread drizzle, and the surface temperature warms.

Fig. 12.32 shows how the static stability relates to the type of occlusion. In this sketch of a warm front occlusion, look at the static stability across the occluded front, at the altitude of the dashed line. To the east of the front in the cold airmass, the air is strongly stable (as shown by the tight packing of the isentropes in the vertical). To the west in the cool airmass, the air is less statically stable (i.e., greater vertical spacing between isentropes). If this had been a cold frontal occlusion instead, the greatest static stability would have been west of the front.

The wedge of warm air (Fig. 12.32) pushed up between the cool and cold airmasses is called a **TROWAL**, an acronym that means "**trough of warm air aloft**." This TROWAL, labeled in Fig. 12.31a, touches the ground at the triple point, but tilts toward higher altitudes further north. Under the TROWAL can be significant precipitation and severe weather at the surface (Fig. 12.31b) — hence it is important for weather forecasting.

The previous sketches are "text-book" examples of prototypical situations. In real life, more complex maps and cross sections can occur. Sometimes more than three airmasses can be drawn together in a low, causing multiple cold or warm fronts, and multiple fronts aloft.

12.7. BENT-BACK FRONTS & STING JETS

Fronts are intimately tied to mid-latitude cyclones. One approximation of this connection is called the **Norwegian cyclone model**, where the cold and warm fronts intersect (Figs. 12.30a & 12.31a) or where they pivot around a common low-pressure center (Fig. 12.1). The next chapter discussed extratropical cyclones in more detail.

Another model, which has been found to describe some marine cyclones, is called the **Shapiro-Keyser cyclone model** (Fig. 12.33). In this model, the cold front is roughly perpendicular to, but does not intersect, the warm front.

Sometimes the warm/occluded front wraps around the low-center. The result is a **bent-back front** (Fig. 12.33) equatorward of the low. This portion of the front has cold air advancing poleward, and is identified as a weakening cold front.

Near the end of some bent-back fronts is a curved region of fast mid-tropospheric air that descends to the Earth's surface causing a destructive windstorm called a **sting jet**. Its name is motivated by the curved poisonous tail (stinger) of a scorpion.

12.8. UPPER-TROPOSPHERIC FRONTS

Upper-tropospheric fronts are also called **upper-level fronts**, and are sometimes associated with **folds in the tropopause**. A cross-section through an idealized upper-level front is sketched in Fig. 12.34.

In the lower troposphere, the south-to-north temperature gradient creates the jet stream due to the thermal wind relationship (see the General Circulation chapter). A reversal of the meridional temperature gradient above 10 km in this idealized sketch causes wind velocities to decrease above that altitude. Within the stratosphere, the isentropes are spaced closer together, indicating greater static stability.

South of the core of the jet stream, the tropopause is relatively high. To the north, it is lower. Between these two extremes, the tropopause can wrap around the jet core, and fold back under the jet, as sketched in Fig. 12.34. Within this fold the isentropes are tightly packed, indicating an **upper-tropospheric front**. Sometimes the upper-level front connects with a surface front (Fig. 12.19).

Stratospheric air has unique characteristics that allow it to be traced. Relative to the troposphere, stratospheric air has high ozone content, high radioactivity (due to former nuclear bomb testing), high static stability, low water-vapor mixing ratio, and high **isentropic potential vorticity**.

The tropopause fold brings air of stratospheric origin down into the troposphere. This causes an injection of ozone and radioactivity into the troposphere. The dry air in the tropopause fold behind the cold front is often clearly visible in water-vapor satellite images of the upper troposphere.

The heavy line in Fig. 12.34 corresponds roughly to the 1.5 PVU (potential vorticity units) isopleth,

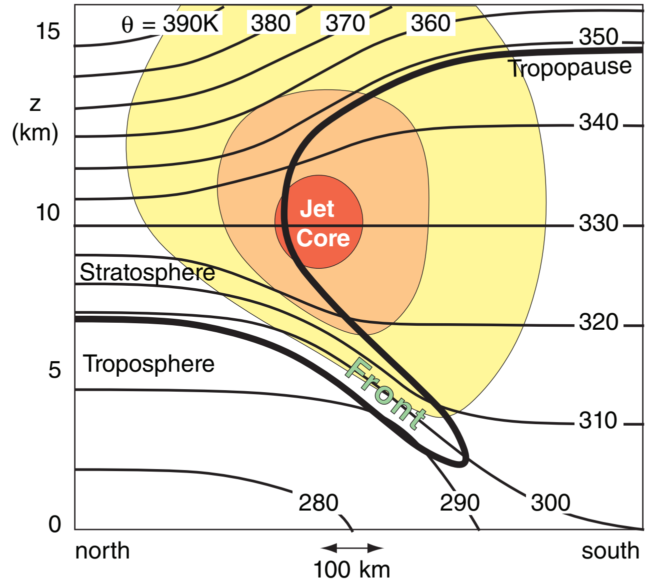


Figure 12.34
Idealized vertical cross-section through the jet stream in N. Hemisphere. Shading indicates west-to-east speeds (into the page): yellow ≥ 50 m s^{-1}, tan ≥ 75 m s^{-1}, orange ≥ 100 m s^{-1}. Thin lines are isentropes (K). Thick line is the tropopause. "Front" indicates the upper-tropospheric front, where stratospheric air is penetrating into the troposphere.

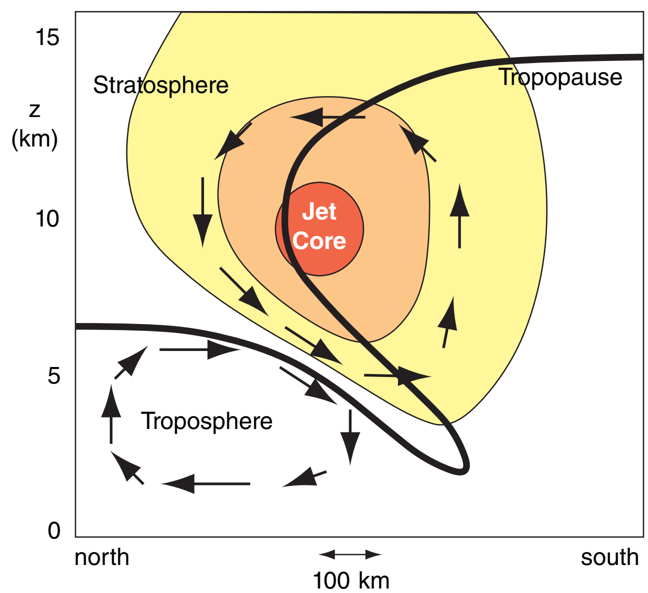


Figure 12.35
Transverse (cross-jet-stream) circulations (arrows) that dynamically form an upper-tropospheric front (i.e., a tropopause fold).

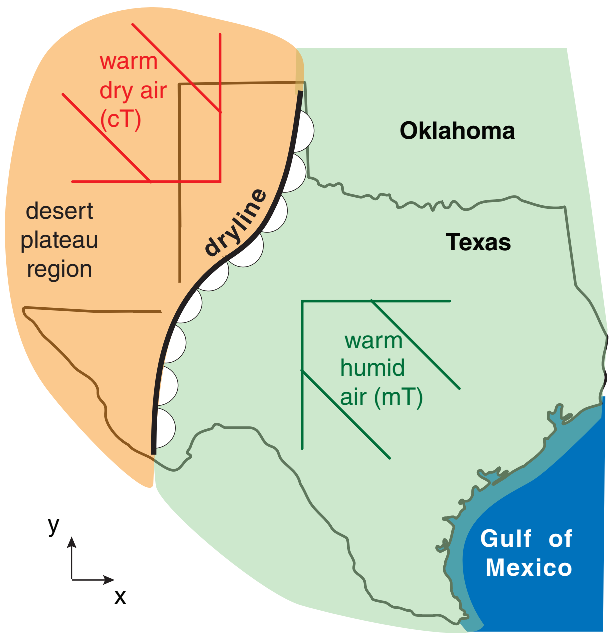


Figure 12.36
Typical location of dryline in the southwestern USA.

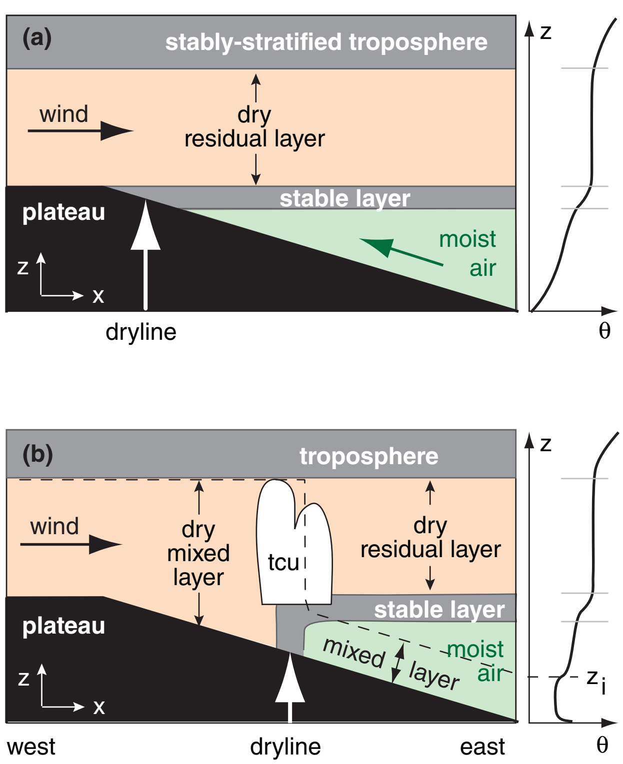


Figure 12.37
Idealized vertical cross section through a dryline (indicated by white arrow). Winds are indicated with black arrows. (a) Early morning. (b) Mid afternoon, with a convective mixed layer of thickness z_i. tcu = towering cumulus clouds. At right of each figure is the potential temperature θ sounding for the east side of the domain.

and is a good indicator of the tropopause. Above this line, PVU values are greater than 1.5, and increase rapidly with height to values of roughly 10 PVU at the top of Fig. 12.34. Below this line, the PVU gradient is weak, with typical values of about 0.5 PVU at mid-latitudes. PVU is negative in the Southern Hemisphere, but of similar magnitude.

Sawyer-Eliassen dynamics help create tropopause folds. Picture a tropopause that changes depth north to south, as in Figs. 11.35 or 11.37 of the General Circulation chapter, but does not yet have a fold. If the jet core advects colder air into the region, then the thickness between the 10 and 20 kPa isobaric surfaces (Fig. 11.37b) will decrease as described by the hypsometric equation. This will cause greater slope (i.e., greater height gradient) of the 10 kPa surface near the jet core. This temporarily upsets geostrophic balance, causing an ageostrophic circulation above the jet from south to north.

This transient flow continues to develop into a cross jet-stream direct circulation that forms around the jet as sketched in Fig. 12.35 (similar to Fig. 12.28b). Below the fold is an indirect circulation that also forms due to a complex geostrophic imbalance. Both circulations distort the tropopause to produce the fold, as sketched in Fig. 12.35, and strengthen the upper-level front due to the kinematic tilting term.

12.9. DRYLINES

In western Texas, USA, there often exists a boundary between warm humid air to the east, and warm dry air to the west (Fig. 12.36). This boundary is called the **dryline**. Because the air is hot on both sides of the boundary, it cannot be labeled as a warm or cold front. The map symbol for a dryline is like a warm front, except with open semicircles adjacent to each other, pointing toward the moist air.

Drylines are observed during roughly 40 to 50% of all days in spring and summer in that part of the world. Similar drylines are observed in other parts of the world.

During midday through afternoon, convective clouds are often triggered along the drylines as the less-dense moist air rides over the denser dry air. Some of these cloud bands grow into organized thunderstorm squall lines that can propagate east from the dryline.

Drylines need not be associated with a wind shift, convergence, vorticity, nor with low pressure. Hence, they do not satisfy the definition of a front. However, sometimes drylines combine with troughs to dynamically contribute to cyclone development.

Drylines in the southwestern USA tend to move eastward during the morning, and return westward during evening. This diurnal cycle is associated with the daily development of a convective mixed layer, interacting with the sloping terrain, via the following mechanism.

Between the Gulf of Mexico (at sea level) and the high-desert plateau of west Texas and New Mexico (roughly 1.5 km above sea level) is a broad region of sloping terrain. Over the high plateaus to the west, deep hot turbulent mixed layers can form in the late afternoon. By night and early morning, after turbulence decays, this thick dry layer of hot air over the plateau becomes a non-turbulent residual layer with nearly adiabatic lapse rate (Fig. 12.37a). Westerly winds advect this dry air eastward, where it overrides cooler, humid air at lower altitudes. Between that elevated residual layer and the moister air below is a strong stable layer such as a temperature inversion.

After sunrise, the warm ground causes a convective mixed-layer to form and grow. The mixed-layer depth is shown by the dashed line in Fig. 12.37b. Because the ground is higher in the west, the top of the mixed layer is able to reach the dry residual-layer air in the west earlier than in the east. When this dry air is entrained downward into the air below, it causes the surface humidity to drop.

Meanwhile, further east, the top of the mixed layer has yet to reach the dry air aloft, so the humidity is still high there. The dryline separates this moist surface air in the east from the dry air to the west.

As the day continues, the mixed layer in the moist air deepens, allowing the region of dry-air entrainment to spread eastward. This causes the dryline to move eastward as the day progresses. At night, convective turbulence ceases, a stable boundary layer develops, and prevailing low-altitude easterlies advect moist air back toward the west (Fig. 12.37a).

If we assume that initially the inversion is level with respect to sea level, then using geometry, we can see that the progression speed of the dryline eastward $\Delta x/\Delta t$ is directly related to the rate of growth of the mixed layer $\Delta z_i/\Delta t$:

$$\frac{\Delta x}{\Delta t} = \frac{\Delta z_i \,/\, \Delta t}{s} \tag{12.13}$$

where $s = \Delta z/\Delta x$ is the terrain slope.

Suppose the early morning stable boundary layer has a linear profile of potential temperature $\gamma = \Delta\theta/\Delta z$, then the mixed-layer growth proceeds with the square-root of accumulated daytime heating Q_{Ak} (see the Atmospheric Boundary Layer chapter):

$$z_i = \left(\frac{2 \cdot Q_{Ak}}{\gamma}\right)^{1/2} \tag{12.14}$$

Sample Application (§)

Suppose the surface heat flux during the day is constant with time at 0.2 K·m s^{-1} in the desert southwest USA. The early morning sounding has a potential temperature gradient of γ = 10 K km^{-1}. The terrain slope is 1:400. Plot the dryline position with time.

Find the Answer

Given: γ = 10 K km^{-1}, s = 1/400, F_H = 0.2 K·m s^{-1},
Find: $\Delta x(t)$ = ? km.

For constant surface flux, cumulative heating is:

$$Q_{Ak} = F_H \cdot \Delta t$$

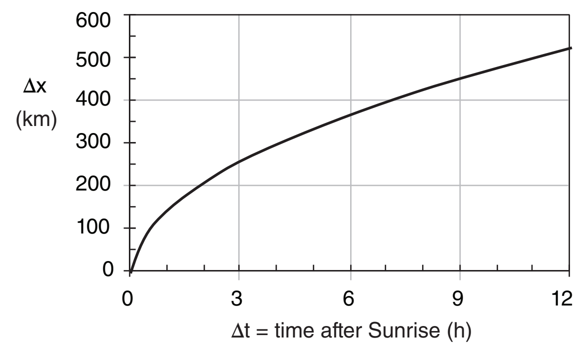
where Δt is time since sunrise.

Thus, eq. (12.15) can be rewritten as:

$$\Delta x = \frac{1}{s} \cdot \left(\frac{2 \cdot F_H \cdot \Delta t}{\gamma}\right)^{1/2}$$

$$= \frac{400}{1} \cdot \left[\frac{2 \cdot (0.2\mathrm{K \cdot m/s}) \cdot \Delta t}{(0.01\mathrm{K/m})}\right]^{1/2}$$

Plotting this with a spreadsheet program:



Check: Units OK. Physics OK. Graph reasonable.
Exposition: The dryline moves to the east with the square-root of time. Thus, the further east it gets, the slower it goes.

As a result, the distance Δx that a dryline moves eastward as a function of time is:

$$\Delta x = \frac{1}{s} \cdot \left(\frac{2 \cdot Q_{Ak}}{\gamma} \right)^{1/2} \qquad (12.15)$$

Time is hidden in the cumulative heat term Q_{Ak}.

12.10. REVIEW

Surface fronts mark the boundaries between airmasses. Changes in wind, temperature, humidity, visibility, and other meteorological variables are frequently found at fronts. Along frontal zones are often low clouds, low pressure, and precipitation. Fronts rotate counterclockwise around the lows in the Northern Hemisphere as the lows move and evolve.

Fronts strengthen or weaken due to advection by the wind (kinematics), external diabatic heating (thermodynamics), and ageostrophic cross-frontal circulations (dynamics). One measure of frontal strength is the temperature change across it. The spread of cold air behind a front is constrained by Coriolis force, as winds adjust toward geostrophic.

Airmasses can form when air resides over a surface sufficiently long. This usually happens in high-pressure centers of light wind. Airmasses over cold surfaces take longer to form than those over warmer surfaces because turbulence is weaker in the statically stable air over the cold surface. As airmasses move from their source regions, they are modified by the surfaces and terrain over which they flow.

Upper-tropospheric fronts can form as folds in the tropopause. Drylines form over sloping terrain as a boundary between dry air to the west and moist air to the east; however, drylines do not usually possess all the characteristics of fronts.

12.11. HOMEWORK EXERCISES

12.11.1. Broaden Knowledge & Comprehension

B1. Monitor the weather maps on the web every day for a week or more during N. Hemisphere winter, and make a sketch of each low center with the fronts extending (similar to Fig. 12.1). Discuss the variety of arrangements of warm, cold, and occluded fronts that you have observed during that week.

B2. Same as B1, but for the S. Hemisphere during S. Hemisphere winter.

B3. On a surface weather map, identify high-pressure centers, and identify ridges. Also, on a 50 kPa (500 hPa) chart, do the same. Look at the ratio of ridges to high centers for each chart, and identify which chart (surface or 50 kPa) has the largest relative number of high centers, and which has the largest relative number of ridges.

B4. From weather maps showing vertical velocity w, find typical values of that vertical velocity near the center of highs. (If $\omega = \Delta P/\Delta t$ is used instead of w, it is acceptable to leave the units in mb s^{-1}). Compare this vertical velocity to the radial velocity of air diverging away from the high center in the boundary layer.

B5. Use a sequence of surface weather maps every 6 or 12 h for a week, where each map spanning a large portion of the globe. Find one or more locations that exhibits the following mechanism for formation of a surface high pressure:

a. global circulation b. monsoon
c. Rossby wave d. thunderstorm
e. topographic

B6. Do a web search to find surface weather maps that also include airmass abbreviations on them. Print one of these maps, and suggest how the labeled airmass moved to its present location from its genesis region (where you will need to make a reasonable guess as to where its genesis region was).

B7. Use weather maps on the web to find a location where air has moved over a region and then becomes stationary. Monitor the development of a new airmass (or equivalently, modification of the old airmass) with time at this location. Look at temperature and humidity.

B8. For a weather situation of relatively zonal flow from west to east over western N. America, access upper air soundings for weather stations in a line more-or-less along the wind direction. Show how the sounding evolves as the air flows over each major mountain range. How does this relate to airmass modification?

B9. Access upper air soundings for a line of RAOB stations across a front. Use this data to draw vertical cross-sections of:
a. pressure or height b. temperature
c. humidity d. potential temperature

B10. Same as previous exercise, but identify the frontal inversions (or frontal stable layers) aloft in the soundings.

B11. Access a sequence of weather maps that cover a large spatial area, with temporal coverage every 6 or 12 h for a week. Find an example of a front that you can follow from beginning (frontogenesis) through maturity to the end (frontolysis). Discuss how the front moves as it evolves, and what the time scale for its evolution is.

B12. Print from the web a weather map that shows both the analyzed fronts and the station plot data. On this map, draw your own analysis of the data to show where the frontal zones are. Discuss how the weather characteristics (wind, pressure, temperature, weather) across these real fronts compare with the idealized sketches of Figs. 12.11 and 12.12.

B13. Access a sequence of surface weather maps from the web that show the movement of fronts. Do you see any fronts that are labeled as stationary fronts, but which are moving? Do you see any fronts labeled as warm or cold fronts, but which are not moving? Are there any fronts that move backwards compared to the symbology labeling the front (i.e., are the fronts moving opposite to the direction that the triangles or semicircles point)? [Hint: often fronts are designated by how they move relative to the low center. If a cold front, for example, is advancing cyclonically around a low, but the low is moving toward the west (i.e., backwards in mid-latitudes), then relative to people on the ground, the cold front is retreating.]

B14. Access weather map data that shows a strong cold front, and compare the winds on both sides of that front to the winds that you would expect using geostrophic-adjustment arguments. Discuss.

B15. For a wintertime situation, access a N. (or S.) Hemisphere surface weather map from the web, or access a series of weather maps from different agencies around the world in order to get information for the whole Hemisphere. Draw the location of the polar front around the globe.

B16. Access weather maps from the web that show a strengthening cold or warm front. Use the other weather data on this map to suggest if the strengthening is due mostly to kinematic, thermodynamic, or dynamic effects.

B17. Access a sequence of weather maps that shows the formation and evolution of an occluded front. Determine if it is a warm or cold occlusion (you might need to analyze the weather map data by hand to help you determine this).

B18. Use the web to access a 3-D sketch of a TROWAL. Use this and other web sites to determine other characteristics of TROWALs.

B19. Download upper-air soundings from a station under or near the jet stream. Use the data to see if there is a tropopause fold. [Hint: assuming that you don't have measurements of radioactivity or isentropic potential vorticity, use mixing ratio or potential temperature as a tracer of stratospheric air.]

B20. Search the web for surface weather maps that indicate a dryline in the S.W. USA. If one exists, then search the web for upper-air soundings just east and just west of the dryline. Plot the resulting soundings of potential temperature and humidity, and discuss how it relates to the idealized sketch of a dryline in this textbook.

12.11.2. Apply

A1. Identify typical characteristics of the following airmass:
a. cAA b. cP c. cT d. cM e. cE
f. mA g. mP h. mT i. mM j. mE
k. cEw l. cAk m. cPw n. mPk o. mTw

A2. List all the locations in the world where the following airmasses typically form.
a. cAA b. cA c. cP d. mP e. mT
f. cT g. mE

A3.(§) Produce a graph of warm airmass depth z_i and airmass potential temperature θ_{ML} as a function of time, similar to Fig. 12.5. Use all the same conditions as in that figure (γ = 3.3 K km^{-1}, $\beta = 10^{-6}$ s^{-1}, initial $\Delta\theta s$ = 10°C; see the Sample Application on the subsequent page) for genesis of a warm airmass, except with the following changes:
a. initial $\Delta\theta s$ = 5°C b. initial $\Delta\theta s$ = 15°C
c. initial $\Delta\theta s$ = 20°C d. initial $\Delta\theta s$ = 15°C

e. $\beta = 0\ s^{-1}$ f. $\beta = 10^{-7}\ s^{-1}$
g. $\beta = 5 \times 10^{-7}\ s^{-1}$ h. $\beta = 5 \times 10^{-6}\ s^{-1}$
i. $\beta = 10^{-5}\ s^{-1}$ j. $\gamma = 5\ K\ km^{-1}$
k. $\gamma = 4\ K\ km^{-1}$ l. $\gamma = 3\ K\ km^{-1}$
m. $\gamma = 2\ K\ km^{-1}$ n. $\gamma = 1\ K\ km^{-1}$

A4. Same as the previous exercise, but find the time scale τ that estimates when the peak airmass depth occurs for warm airmass genesis.

A5. For the katabatic winds of Antarctica, find the downslope buoyancy force per unit mass for this location:

Location	$\Delta z/\Delta x$	$\Delta\theta$ (K)	T_e (K)
a. Interior	0.001	40	233
b. Interior	0.002	35	238
c. Interior	0.003	30	243
d. Intermediate	0.004	27	245
e. Intermediate	0.005	25	248
f. Intermediate	0.006	24	249
g. Intermediate	0.007	23	250
h. Coast	0.008	22	251
i. Coast	0.009	21	252
j. Coast	0.010	20	253
k. Coast	0.011	19	254

A6.(§) An Arctic airmass of initial temperature –30°C is modified as it moves at speed 10 m s^{-1} over smooth warmer surface of temperature 0°C. Assume constant airmass thickness. Find and plot the airmass (mixed-layer) temperature vs. downwind distance for an airmass thickness (m) of:
a. 100 b. 200 c. 300 d. 400 e. 500
f. 600 g. 700 h. 800 i. 900 j. 1000
k. 1200 l. 1400 m. 1500

A7. Find the external Rossby radius of deformation at 60° latitude for a cold airmass of thickness 500 m and $\Delta\theta$ (°C) of:
a. 2 b. 4 c. 6. d. 8 e. 10 f. 12 g. 14
h. 16 i. 18 j. 20 k. 22 l. 24 m. 26 n. 28
Assume a background potential temperature of 300K.

A8.(§) Find and plot the airmass depth and geostrophic wind as a function of distance from the front for the cases of the previous exercise. Assume a background potential temperature of 300 K.

A9. Suppose that $\Delta\theta/\Delta x$ = 0.02°C km^{-1}, $\Delta\theta/\Delta y$ = 0.01°C km^{-1}, and $\Delta\theta/\Delta z$ = 5°C km^{-1}. Also suppose that $\Delta U/\Delta x$ = – 0.03 (m/s) km^{-1}, $\Delta V/\Delta x$ = 0.05 (m/s) km^{-1}, and $\Delta W/\Delta x$ = 0.02 (cm/s) km^{-1}. Find the kinematic frontogenesis contributions from:
a. confluence b. shear c. tilting
d. and find the strengthening rate.

A10. A thunderstorm on the warm side of a 300 km wide front rains at the following rate (mm h^{-1}). Find the thermodynamic contribution to frontogenesis.
a. 0.2 b. 0.4 c. 0.6 d. 0.8 e. 1.0 f. 1.2 g. 1.4
h. 1.6 i. 1.8 j. 2.0 k. 2.5 l. 3.0 m. 3.5 n. 4.0

A11. Plot dryline movement with time, given the following conditions. Surface heat flux is constant with time at kinematic rate 0.2 K·m s^{-1}. The vertical gradient of potential temperature in the initial sounding is γ. Terrain slope is $s = \Delta z/\Delta x$.

	γ (K km^{-1})	s		γ (K km^{-1})	s
a.	8	1/500	b.	8	1/400
c.	8	1/300	d.	8	1/200
e.	12	1/500	f.	12	1/400
g.	12	1/300	h.	12	1/200
i.	15	1/500	j.	15	1/400

12.11.3. Evaluate & Analyze

E1. Would you expect there to be a physically- or dynamically-based upper limit on the number of warm fronts and cold fronts that can extend from a low-pressure center? Why?

E2. Sketch a low with two fronts similar to Fig. 12.1, but for the Southern Hemisphere.

E3. What are the similarities and differences between a ridge and a high-pressure center? Also, on a weather map, what characteristics would you look for to determine if a weather feature is a ridge or a trough?

E4. In high-pressure centers, is the boundary-layer air compressed to greater density as subsidence pushes down on the top of the mixed layer? Explain.

E5. In Fig. 12.3b, use the hypsometric relationship to explain why the ridge shifts or tilts westward with increasing altitude. Hint: consider where warm and cold air is relative to the surface high.

E6. How would Fig. 12.3b be different, if at all, in Southern Hemisphere mid-latitudes. Hint: Consider the direction that air rotates around a surface high, and use that information to describe which side of the high (east or west) advects in warm air and which advects in colder air.

E7. Consider the global-circulation, monsoon, and topographic mechanisms that can form high-pressure regions. Use information from the chapters on General Circulation, Extratropical Cyclones, and Regional Winds to identify 3 or more regions in the world where 2 or more of those mechanisms are superimposed to help create high-pressure regions. Describe the mechanisms at each of those 3 locations, and suggest how high-pressure formation might vary during the course of a year.

E8. Out of all the different attributes of airmasses, why do you think that the airmass abbreviations listed in Table 12-1 focus on only the temperature (i.e., AA, A, P, T, M, E) and humidity (c, m)? In other words, what would make the relative temperature and humidity of airmasses be so important to weather forecasters? Hint: consider what you learned about heat and humidity in earlier chapters.

E9. Fig. 12.4 shows the genesis regions for many different airmasses. At first glance, this map looks cluttered. But look more closely for patterns and describe how you can anticipate where in the world you might expect the genesis regions to be for a particular airmass type.

E10. In this chapter, I described two very different mechanisms by which warm and cold airmasses are created. Why would you expect them to form differently? Hint: consider the concepts, not the detailed equations.

E11. Fig. 12.5 shows how initially-cold air is transformed into a warm airmass after it becomes parked over a warm surface such as a tropical ocean. It is not surprising that the airmass temperature θ_{ML} asymptotically approaches the temperature of the underlying surface. Also not surprising: depth z_i of the changed air initially increases with time.

But more surprising is the decrease in depth of the changed air at longer times. Conceptually, why does this happen? Why would you expect it to happen for most warm airmass genesis?

E12. Using the concept of airmass conservation, formulate a relationship between subsidence velocity at the top of the boundary layer (i.e., the speed that the capping inversion is pushed down in the absence of entrainment) to the radial velocity of air within the boundary layer. (Hint: use cylindrical coordinates, and look at the geometry.)

E13. For the toy model in the INFO Box on Warm Airmass Genesis, ΔQ is the amount of heat (in kinematic units) put into the mixed layer (ML) from the surface during one time step. By repeating over many time steps, we gradually put more and more heat into the ML. This heat is uniformly mixed in the vertical throughout the depth z_i of the ML, causing the airmass to warm.

Yet at any time, such as 10 days into the forecast, the amount of additional heat contained in the ML (which equals its depth times its temperature increase since starting, see Fig. 12.5) is less than the accumulated heat put into the ML from the surface (which is the sum of all the ΔQ from time zero up to 10 days).

That discrepancy is not an error. It describes something physical that is happening. What is the physical process that explains this discrepancy, and how does it work?

E14. In the Sample Application on warm airmass genesis, I iterate with very small time step at first, but later take larger and larger time steps. Why would I need to take small time steps initially, and why can I increase the time step later in the simulation? Also, if I wanted to take small time steps for the whole duration, would that be good or bad? Why?

E15. Suppose that cold air drains katabatically from the center of Antarctica to the edges. Sketch the streamlines (lines that are parallel to the flow direction) that you would anticipate for this air, considering buoyancy, Coriolis force, and turbulent drag.

E16.(§) Assume the katabatic winds of Antarctica result from a balance between the downslope buoyancy force, Coriolis force (assume 70°S latitude), and turbulent drag force at the surface (assume neutral boundary layer with C_D = 0.002). The surface is smooth and the depth of the katabatic layer is 100 m. Neglect entrainment drag at the top of the katabatic flow. The slope, potential temperature difference, and ambient temperature vary according to the table from exercise N5. Find the katabatic wind speed at the locations in the table from exercise N5.

E17. Where else in the world (Fig. 12.8) would you anticipate orographic airmass modification processes similar to those shown in Fig. 12.9? Explain.

E18. For airmass modification (i.e., while an airmass is blowing over a different surface), why do we describe the heat flux from the surface using a bulk transfer relationship ($C_H \cdot M \cdot \Delta\theta$), while for warm airmass genesis of stationary air we used a buoyancy-velocity approach ($b \cdot w_b \cdot \Delta\theta$)?

E19. The Eulerian heat budget from the Heat chapter shows how air temperature change ΔT is related to the heat input over time interval Δt. If this air is also moving at speed M, then during that same time interval, it travels a distance $\Delta x = M \cdot \Delta t$. Explain how you can use this information to get eq. (12.3), and why the wind speed M doesn't appear in that eq.

E20. Look at Fig. 12.9 and the associated Sample Application. In the Sample Application, the air passes twice through the thermodynamic state given by point 1 on that diagram: once when going from point 0 to points 2, and the second time when going from points 2 to point 3.

Similarly, air passes twice through the thermodynamic state at point 3 on the thermo diagram, and twice through point 5.

But air can achieve the same thermodynamic state twice only if certain physical (thermodynamic) conditions are met. What are those conditions, and how might they NOT be met in the real case of air traversing over these mountain ranges?

E21. Background: Recall that a frontal zone separates warmer and cooler airmasses. The warm airmass side of this zone is where the front is drawn on a weather map. This is true for both cold and warm fronts.

Issue: AFTER passage of the cold front is when significant temperature decreases are observed. BEFORE passage of a warm front is when significant warming is observed.

Question: Why does this difference exist (i.e., AFTER vs. BEFORE) for the passage of these two fronts?

E22. Overlay Fig. 12.11 with Fig. 12.12 by aligning the low-pressure centers. Do not rotate the images, but let the warm and cold fronts extend in different directions from the low center. Combine the information from these two figures to create a new, larger figure showing both fronts at the same time, extending from the same low. On separate copies of this merged figure, draw:

a. isotherms b. isobars c. winds d. weather

E23. Suppose you saw from your barometer at home that the pressure was falling. So you suspect that a front is approaching. What other clues can you use (by standing outside and looking at the clouds and weather; NOT by looking at the TV, computer, or other electronics or weather instruments) to determine if the approaching front is a warm or cold front. Discuss, along with possible pitfalls in this method.

E24. Draw new figures (a) - (d) similar to (a) - (d) in Figs. 12.11 and 12.12, but for an occluded front.

E25. Use the columns of temperature data in Fig. 12.13a, and plot each column as a separate sounding on a thermo diagram (See the Atmos. Stability chapter for blank thermo diagrams that you can photocopy for this exercise. Use a skew-T diagram, unless your instructor tells you otherwise). Describe how the frontal zone shows up in the soundings.

E26. Draw isentropes that you might expect in a vertical cross-section through a warm front.

E27. Other than drylines, is it possible to have fronts with no temperature change across them? How would such fronts be classified? How would they behave?

E28. What clouds and weather would you expect with a cold anafront?

E29. Why should the Rossby radius of deformation depend on the depth of the cold airmass in a geostrophic adjustment process?

E30. For geostrophic adjustment, the initial outflow of cold air turns, due to Coriolis force, until it is parallel to the front. Why doesn't it continue turning and point back into the cold air?

E31. For geostrophic adjustment, what is the nature of the final winds, if the starting point is a shallow cylinder of cold air 2 km thick and 500 km radius?

E32. Starting with Fig. 12.18, suppose that ABOVE the bottom contoured surface the temperature is horizontally uniform. Redraw that diagram but with the top two contoured surfaces sloped appropriately for the new temperature state.

E33. By inspection, write a kinematic frontogenesis equation (similar to eq. 12.10), but for an east-west aligned front.

E34. Figs. 12.22 - 12.24 presume that temperature gradients already exist, which can be strengthened by kinematic frontogenesis. Is that presumption valid for Earth's atmosphere? Justify.

E35. For the fronts analyzed in the Sample Application in the section on Surface Fronts – Horizontal Structure, estimate and compare magnitudes of any kinematic, thermodynamic, and dynamic processes that might exist across those fronts, based on the plotted weather data (which includes info on winds, precipitation, etc.). Discuss the relative importance of the various mechanisms for frontogenesis.

E36. Speculate on which is more important for dynamically generating ageostrophic cross-frontal circulations: the initial magnitude of the geostrophic wind or the change of geostrophic wind.

E37. What happens in an occluded front where the two cold air masses (the one advancing behind the cold front, and the one retreating ahead of the warm front) have equal virtual temperature?

E38. Draw isentropes that you expect in a vertical cross-section through a cold front occlusion (where a cold front catches up to a warm front), and discuss the change of static stability across this front.

E39. What types of weather would be expected with an upper-tropospheric front that does not have an associated surface front? [Hint: track movement of air parcels as they ride isentropic surfaces.]

E40. Draw a sketch similar to Fig. 12.35 showing the transverse circulations, but for a vertical cross-section in the Southern Hemisphere.

E41. Would you expect drylines to be possible in parts of the world other than the S.W. USA? If so, where? Justify your arguments.

12.11.4. Synthesize

S1. Fig. 12.2 shows a ridge that is typical of mid-latitudes in the N. Hemisphere. Sketch a ridge for:
  a. Southern Hemisphere mid-latitudes
  b. N. Hemisphere tropics
  c. S. Hemisphere tropics

Hint: Consider the global pressure patterns described in the General Circulation chapter.

S2. What if there was no inversion at the top of the boundary layer? Redraw Fig. 12.3a for this situation.

S3. What if the Earth had no oceans? What mechanisms could create high-pressure centers and/or high-pressure belts?

S4. For your location, rank the importance of the different mechanisms that could create highs, and justify your ranking.

S5. Airmasses are abbreviated mostly by the relative temperature and humidity associated with their formation locations. Table 12-1 also describes two other attributes: returning airmasses, and airmasses that are warmer or colder than the underlying surface. What one additional attribute would you wish airmasses could be identified with, to help you to predict the weather at your location? Explain.

S6. Fig. 12.4 shows many possible genesis regions for airmasses. But some of these regions would not likely exist during certain seasons, because of the absence of high-pressure centers. For the hemisphere (northern or southern) where you live, identify how the various genesis regions appear or disappear with the seasons. Also, indicate the names for any of the high-pressure centers. For those that exist year round, indicate how they shift location with the seasons. (Hint: review the global circulation info in the General Circulation chapter.)

S7. What if airmasses remained stationary over the genesis regions forever? Could there be fronts and weather? Explain.

S8. Would it be possible for an airmass to become so thick that it fills the whole troposphere? If so, explain the conditions needed for this to occur.

S9. What if boundary-layer processes were so slow that airmasses took 5 times longer to form compared to present airmass formation of about 3 to 5 days? How would the weather and global circulation be different, if at all?

S10. Cold-air drainage from Antarctica is so strong and persistent that it affects the global circulation. Discuss how this affect is captured (or not) in the global maps in the General Circulation chapter.

S11. What if Antarctica was flat, similar to the Arctic? How would airmasses and Earth's climate be different, if at all?

S12. Consider Fig. 12.8. How would airmass formation and weather be different, if at all, given:
  a. Suppose the Rocky Mountains disappeared, and a new dominant mountain range (named after you), appeared east-west across the middle of N. America.
  b. Suppose that the Alps and Pyrenees disappeared and were replaced by a new dominant north-south mountain range (named after you) going through Europe from Copenhagen to Rome.
  c. Suppose no major mountain ranges existed.

S13. Suppose a cross-section of the terrain looked like Fig. 12.9, except that the Great Basin region were below sea level (and not flooded with water). Describe how airmass modification by the terrain would be different, if at all.

S14. What if precipitation did not occur as air flows over mountain ranges? Thus, mountains could not modify air masses by this process. How would weather and climate be different from now, if at all?

S15. A really bad assumption was made in eq. (12.3); namely, z_i = constant as the airmass is modified. A better assumption would be to allow z_i to change with time (i.e., with distance as it blows over the new surface). The bad assumption was made because it allowed us to describe the physics with a simple eq. (12.3), which could be solved analytically to get eq. (12.4). This is typical of many physical problems, where it is impossible to find an exact analytical solution to the full equations describing true physics (as best we know it).

So here is a philosophical question. Is it better to approximate the physics to allow an exact analytical solution of the simplified problem? Or is it better to try to get an approximate (iterated or graphical solution) to the exact, more-complicated physics? Weigh the pros and cons of each approach, and discuss.

S16. For a (a) cold front, or a (b) warm front, create station plot data for a weather station 100 km ahead of the front, and for another weather station 100 km behind the front. Hint: use the sketches in Figs. 12.11 and 12.12 to help decide what to plot.

S17. Suppose the width of frontal zones were infinitesimally small, but their lengths remained unchanged. How would weather be different, if at all?

S18. What if turbulence were always so intense that frontal zones were usually 1000 km in width? How would weather be different, if at all?

S19. Does the geostrophic adjustment process affect the propagation distance of cold fronts in the real atmosphere, for cold fronts that are embedded in the cyclonic circulation around a low-pressure center? If this indeed happens, how could you detect it?

S20. The boxes at the top of Figs. 12.22 - 12.24 show the associated sign of key terms for kinematic frontogenesis. If those terms have opposite signs, draw sketches similar to those figures, but showing frontolysis associated with:

a. confluence b. shear c. tilting.

S21. Suppose that condensation of water vapor caused air to cool. How would the thermodynamic mechanism for frontogenesis work for that situation, if at all? How would weather be near fronts, if at all?

S22. Suppose that ageostrophic motions were to experience extremely large drag, and thus would tend to dissipate quickly. How would frontogenesis in Earth's atmosphere be different, if at all?

S23. After cold frontal passage, cold air is moving over ground that is still warm (due to its thermal inertia). Describe how static stability varies with height within the cold airmass 200 km behind the cold front. Draw thermo diagram sketches to illustrate your arguments.

S24. Suppose a major volcanic eruption injected a thick layer of ash and sulfate aerosols at altitude 20 km. Would the altitude of the tropopause adjust to become equal to this height? Discuss.

S25. Suppose the slope of the ground in a dryline situation was not as sketched in Fig. 12.37. Instead, suppose the ground was bowl-shaped, with high plateaus on both the east and west ends. Would drylines exist? If so, what would be their characteristics, and how would they evolve?