POLITECNICO DI MILANO

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Cambiamenti Climatici e Adattamenti negli Ecosistemi e nelle Società

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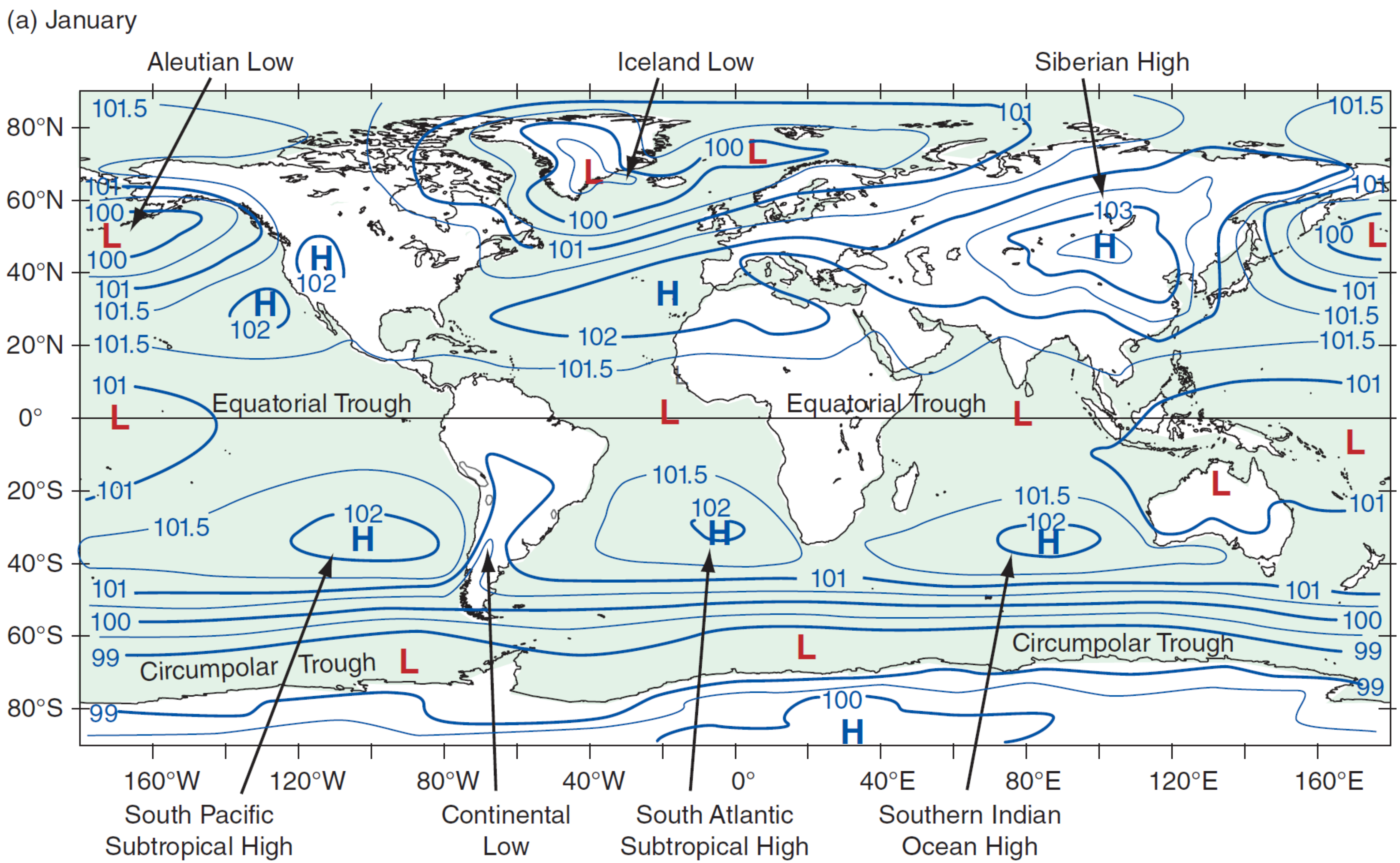
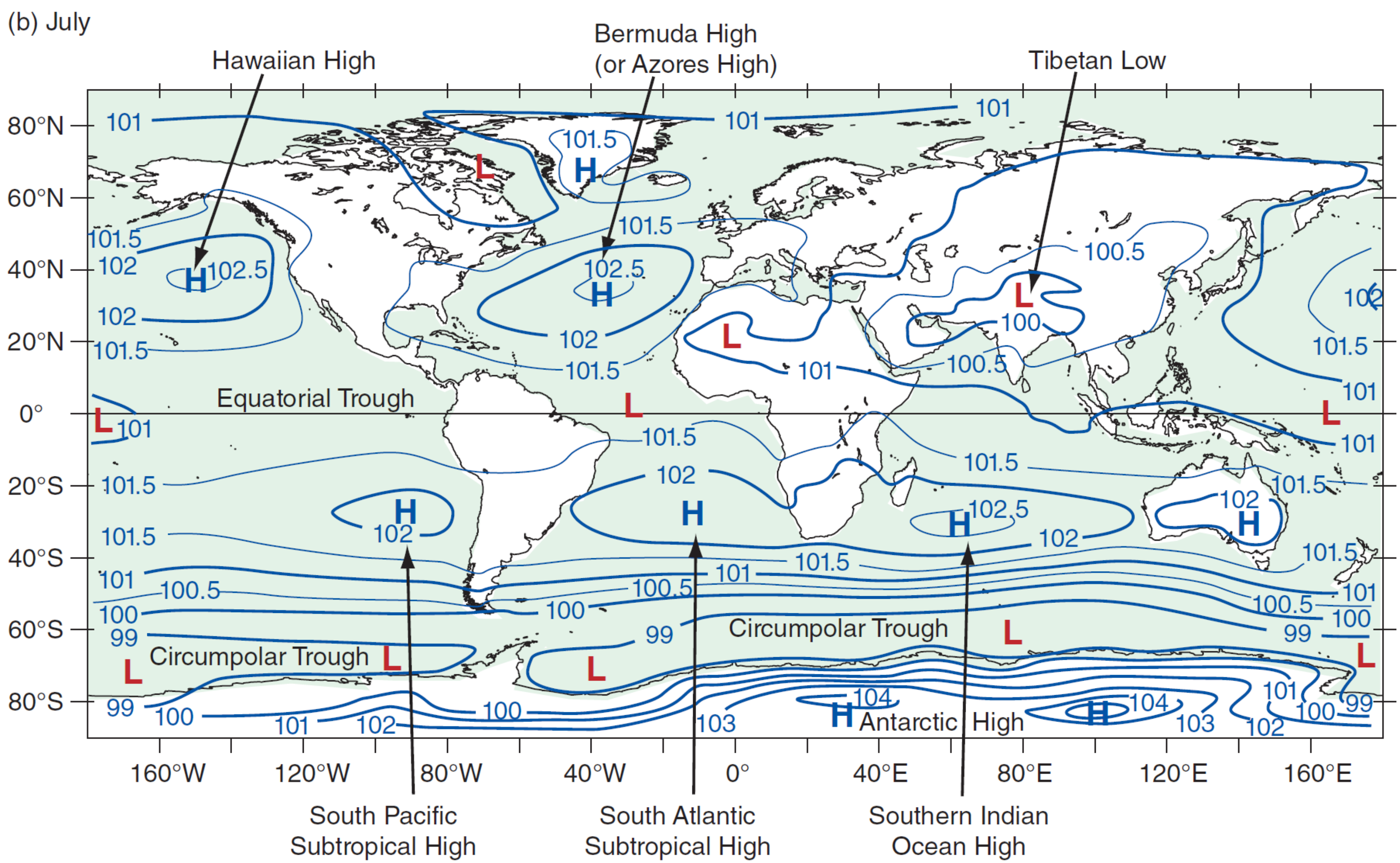
Climate Justice
Jean Monnet
Centre of Excellence

General circulation of the atmosphere



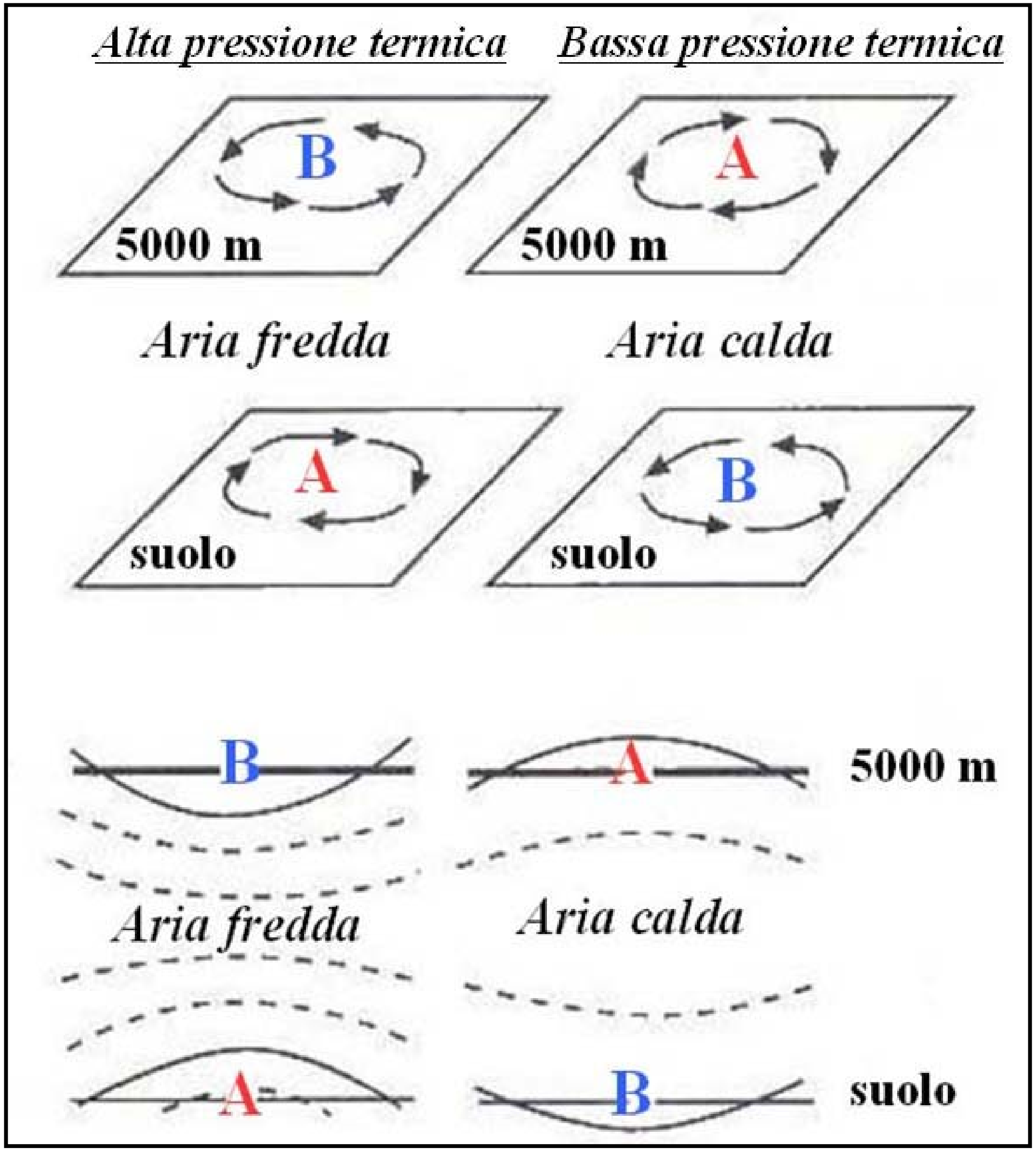
Mean sea-level pressure (kPa) for (a) January, averaged over 30 years: 1981 to 2010.

General circulation of the atmosphere



Mean sea-level pressure (kPa) for (b) July, averaged over 30 years: 1981 to 2010.

Thermal high/low



Thermal high/low

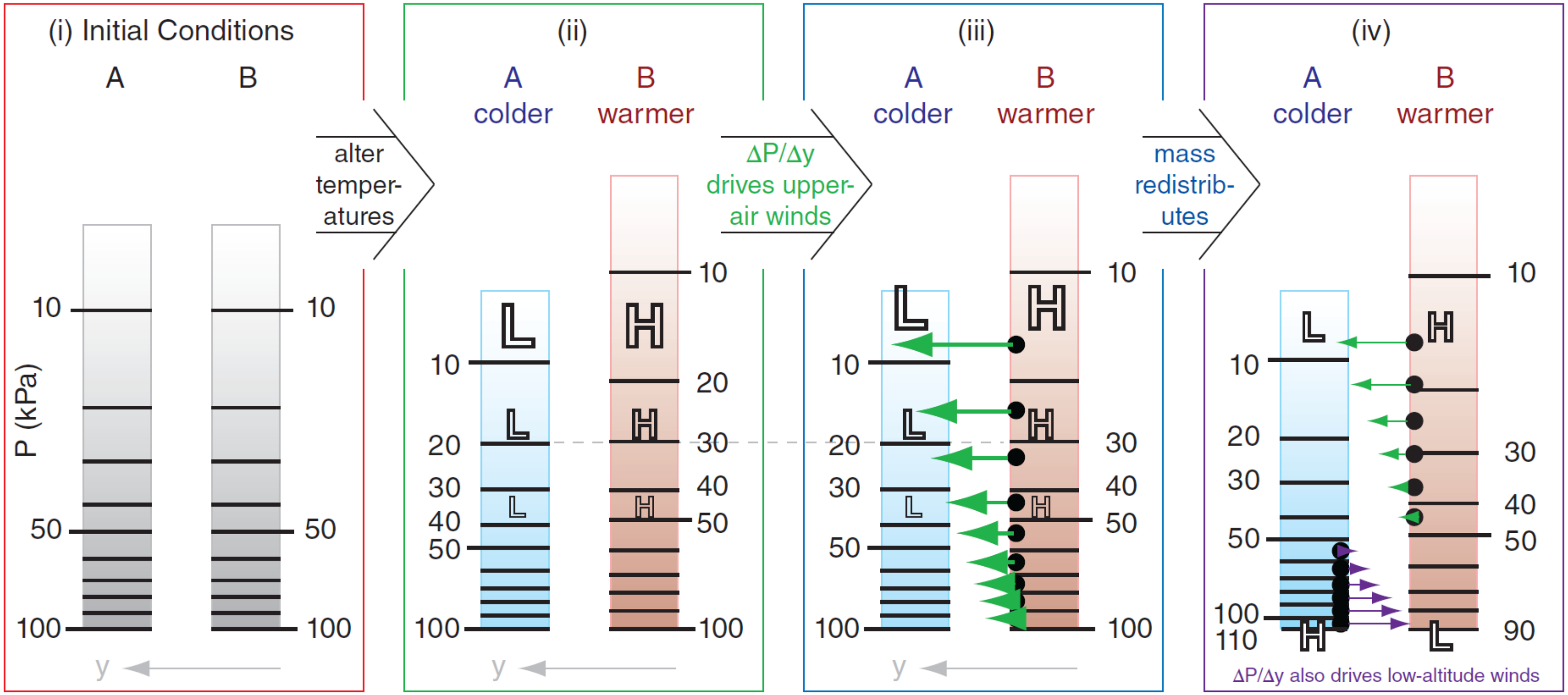


Figure 11.17
Formation of a thermal circulation. The response of two columns of air that are heated differently is that the warmer air column develops a low pressure perturbation at the surface and a high pressure perturbation aloft. Response of the cold column is the opposite. Notation: H = high pressure perturbation (i.e., relative to the average pressure at that altitude). L = low pressure perturbation (relative to the average pressure at that same altitude). Black dots represents air parcels. Thin arrows are winds (i.e., movement of air parcels).

Thermal high/low

Cold columns of air tend to have high surface pressures, while warm columns have low surface pressures.
Next, suppose that some process heats one column relative to the other.
After both columns have finished expanding or contracting due to the temperature change, they will reach new hydrostatic equilibria for their respective temperatures.

The hypsometric equation says that pressure decreases more rapidly with height in cold air than in warm air. Thus, although both columns have the same surface pressure because they contain the same number of molecules, the higher you go above the surface, the greater is the pressure difference between warm and cold air. In Fig. 11.17ii, the printed size of the “H” and “L” indicate the relative magnitudes of the high- and low-pressure perturbations p'.
The horizontal pressure gradient $\Delta P/\Delta y$ aloft between the warm and cold air columns drives horizontal winds from high toward low pressure (Fig. 11.17iii). Since winds are the movement of air molecules, this means that molecules leave the regions of high pressure-perturbation and accumulate in the regions of low. Namely, they leave the warm column, and move into the cold column.
Since there are now more molecules (i.e., more mass) in the cold column, it means that the surface pressure must be greater (H) in the cold column (Fig. 11.17iv). Similarly, mass lost from the warm column results in lower (L) surface pressure. This is called a **thermal low.**

Thermal wind effect

Recall that horizontal temperature gradients cause vertically varying horizontal pressure gradients (Fig. 11.17), and that horizontal pressure gradients drive geostrophic winds. We can combine those concepts to see how horizontal temperature gradients drive vertically varying geostrophic winds. This is called the **thermal wind effect**.

This effect can be pictured via the slopes of isobaric surfaces (Fig. 11.20). The hypsometric equation describes how there is greater thickness between any two isobaric (constant pressure) surfaces in warm air than in cold air. This causes the tilt of the isobaric surfaces to change with altitude. But, tilting isobaric surface imply a pressure-gradient force that can drive the geostrophic wind (*Ug*, *Vg*).The relationship between the horizontal temperature gradient and the changing geostrophic wind with altitude is known as the **thermal wind effect**.

$$\frac{\Delta U_g}{\Delta z} \approx \frac{-|g|}{T_v \cdot f_c} \cdot \frac{\Delta T_v}{\Delta y}$$

$$\frac{\Delta V_g}{\Delta z} \approx \frac{|g|}{T_v \cdot f_c} \cdot \frac{\Delta T_v}{\Delta x}$$

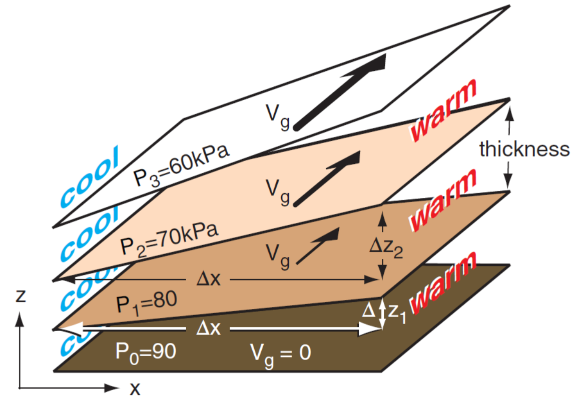
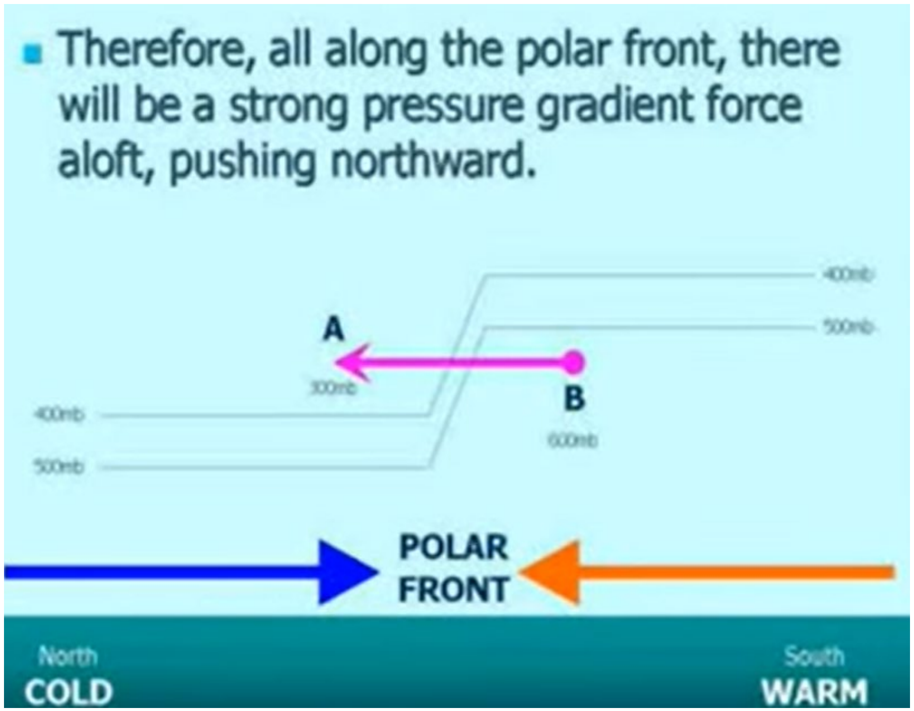
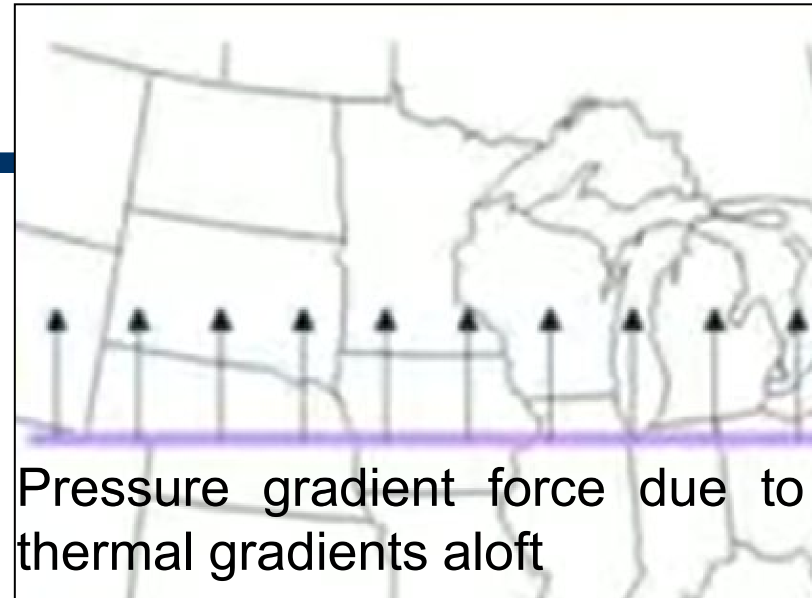
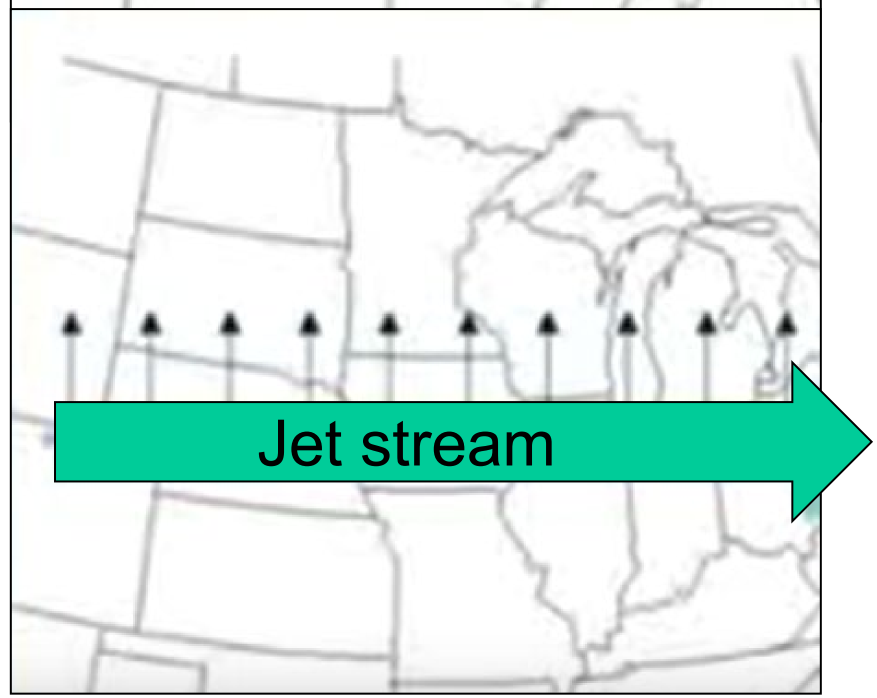


Figure 11.20
Isobaric surfaces are shaded brown. A zonal (west-east) temperature gradient causes the isobaric surfaces to tilt more and more with increasing altitude. Greater tilt causes stronger geostrophic winds (V_g) meridionally (south-north), as plotted with the black vectors for the N. Hemisphere. Geostrophic winds are reversed in S. Hemisphere.

Thermal wind effect







Thermal wind effect

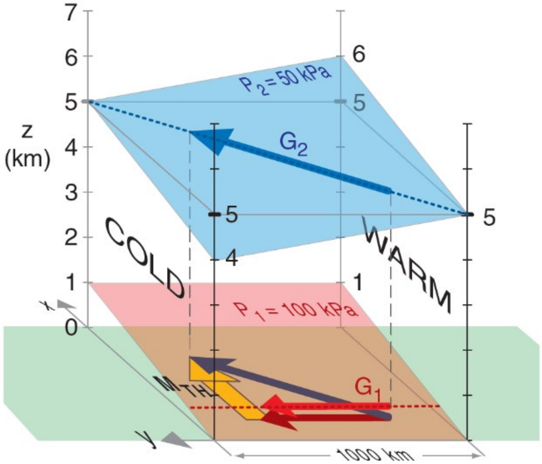
Suppose that we use thickness *TH* as a surrogate for absolute virtual temperature. Then we can combine eqs. (11.14) and (10.29) to yield

$$U_{TH} = U_{G2} - U_{G1} = -\frac{|g|}{f_c}\frac{\Delta TH}{\Delta y}$$

$$V_{TH} = V_{G2} - V_{G1} = +\frac{|g|}{f_c}\frac{\Delta TH}{\Delta x}$$

where U_{TH} and V_{TH} are **components of the thermal wind**, $|g|$ = magnitude of gravitational-acceleration, fc = Coriolis parameter, (U_{G1}, V_{G1}) are geostrophic-wind components on the P_1 isobaric surface, and (U_{G2}, V_{G2}) are geostrophic-wind components on the P_2 isobaric surface.

The horizontal vector defined by (U_{TH}, V_{TH}) is the difference between the geostrophic wind vector on the P_2 surface and the geostrophic wind vector on the P_1 surface, as Figure demonstrates. The corresponding **magnitude of the thermal wind** M_{TH} is:



$$M_{TH} = \sqrt{U_{TH}^{\,2} + V_{TH}^{\,2}}$$

Jet stream

In the winter hemisphere there are often two strong jet streams of fast west-to-east moving air near the tropopause: the **polar jet stream** and the **subtropical jet stream**

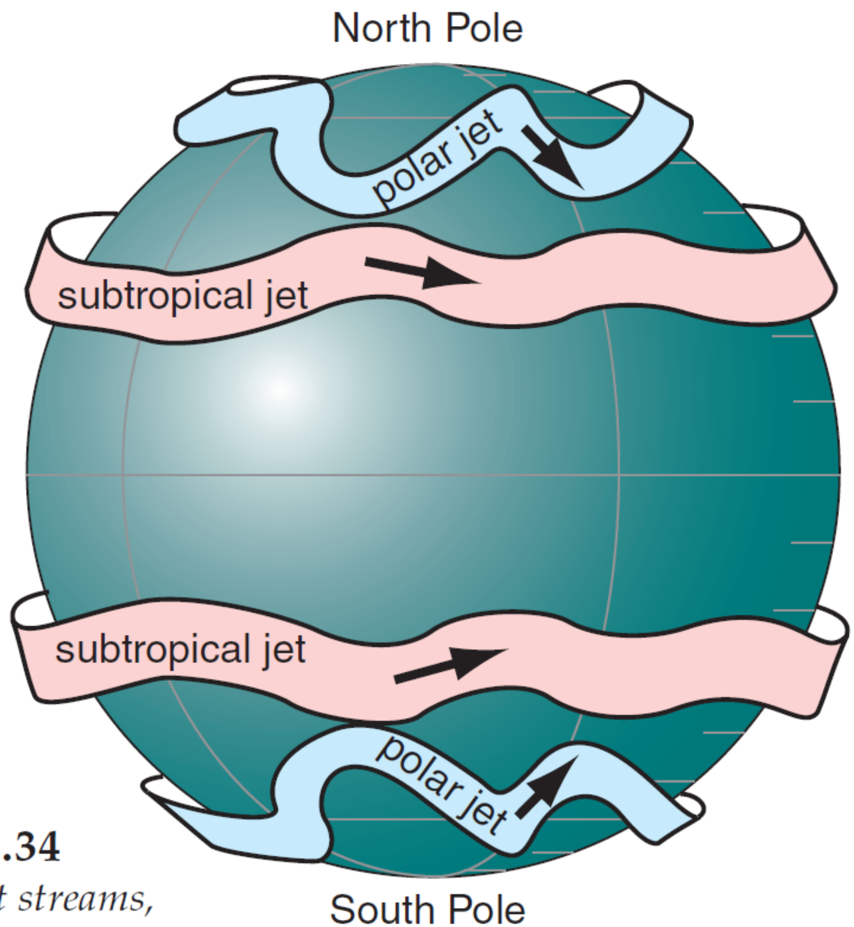
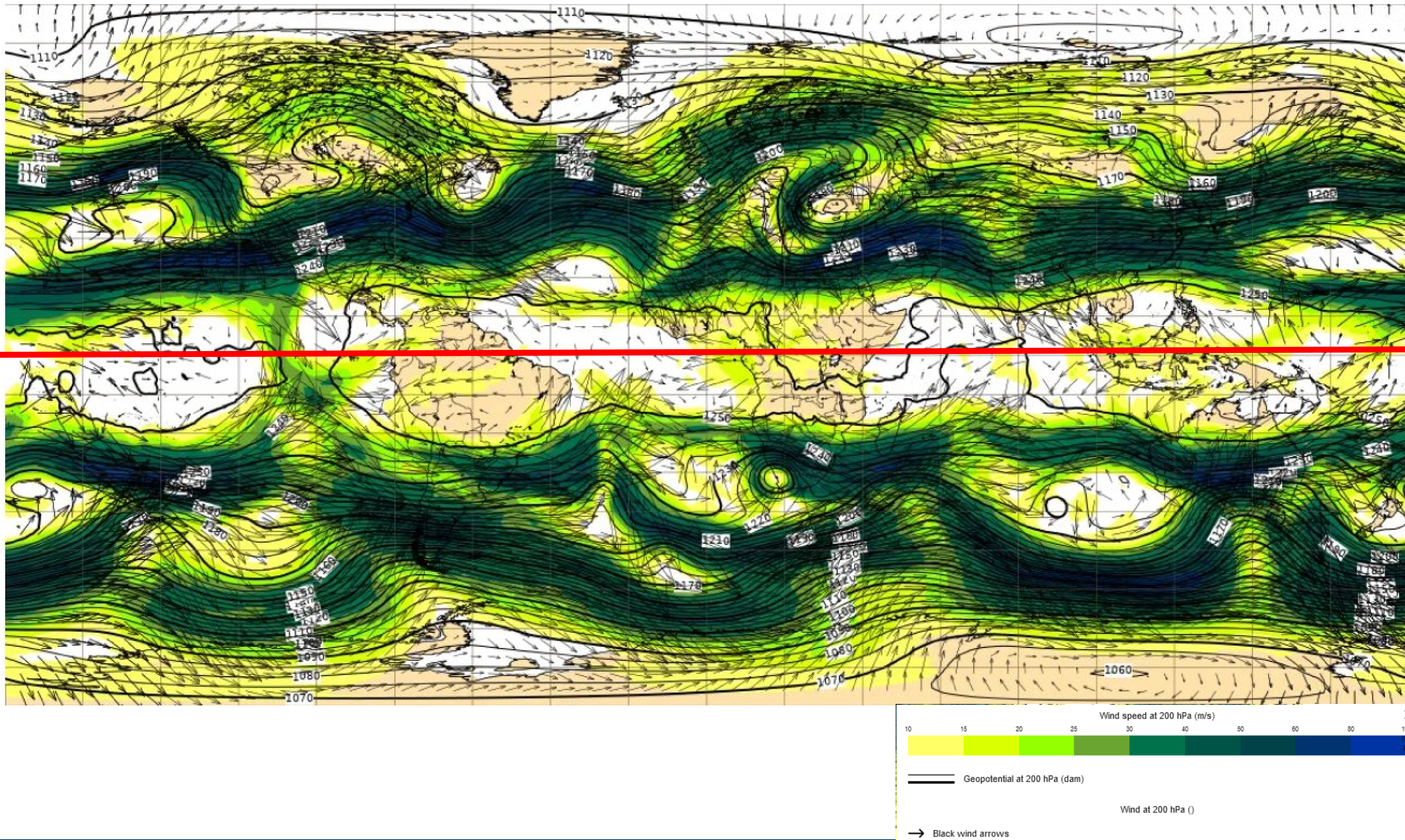


Figure 11.34
Sketch of jet streams, representing a snapshot.

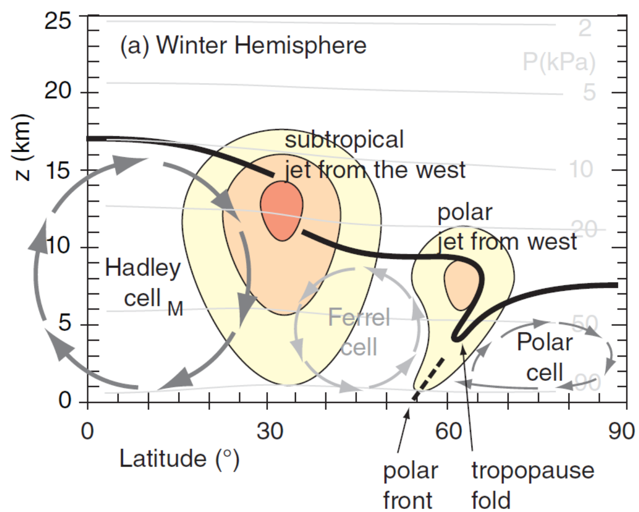
Jet stream (today)



Jet stream in winter

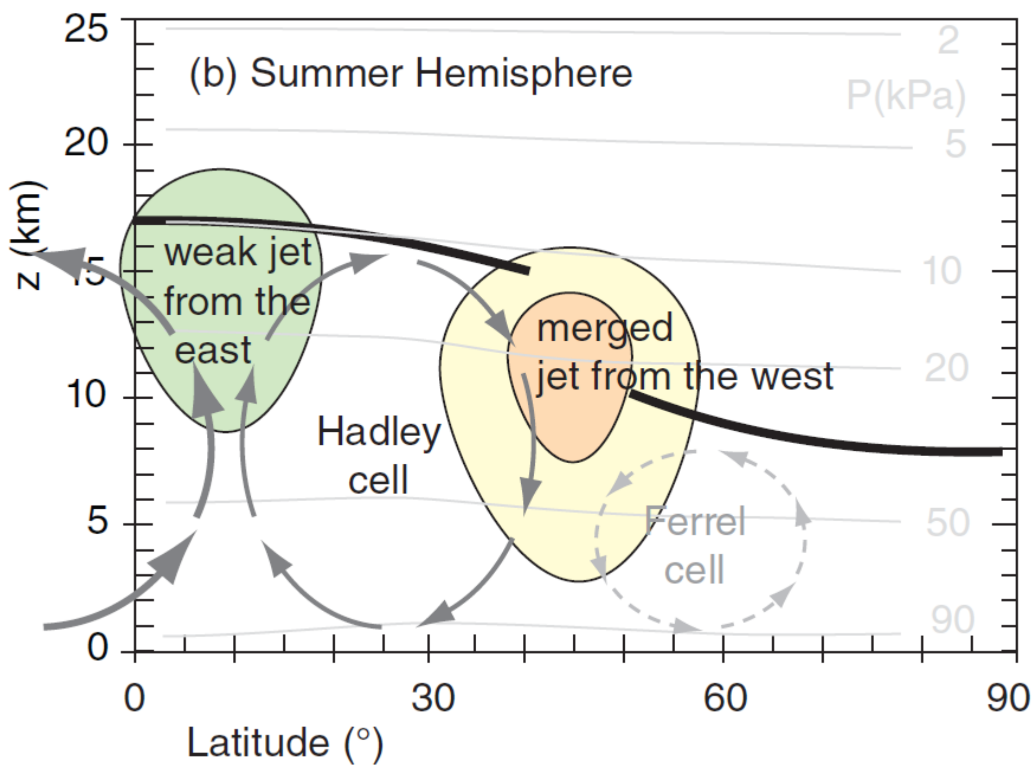
The **subtropical jet** is centered near 30° latitude in the winter hemisphere. This jet: (1) is very steady; (2) meanders north and south a bit; (3) is about 10° latitude wide (width ≈1,000 km); and (4) has seasonal-average speeds of about 45/80 m/s (160/290 km/h). The **core** of fast winds near its center is at 12 km altitude (Figure). It is driven by outflow from the top of the Hadley cell, and is affected by both Coriolis force and angular-momentum conservation.

The **polar jet** is centered near 50 to 60° latitude in the winter hemisphere. The polar jet: (1) is extremely variable; (2) meanders extensively north and south; (3) is about 5° latitude wide; and (4) has widely varying speeds (25 to 100 m/s – 90 to 360 km/h) driven by varying horizontal temperature gradients. The core altitude is about 9 km.



Jet stream in summer

In the summer hemisphere, the jets from the west have merged (Figure), and the winds are slower because of the weaker temperature contrast between the equator and the warm pole. Core wind speeds in the jet are 0 to 10 m/s (0-36 km/h) in N. Hemisphere summer, and 5 to 45 m/s (20-160 km/h) in S. Hemisphere summer. This core shifts poleward to be centered near 40° to 45° latitude. A weak jet from the east is centered at roughly 10° latitude in the summer hemisphere.



Extratropical ridges & troughs (Rossby waves)

To first order, we would expect this jet stream to encircle the globe (Fig. 11.49a) along the zone between the warm and cool air masses, at roughly 50 to 60° latitude in winter. However, this flow is unstable, allowing small disturbances (e.g., flow over mountain ranges) to grow into large north-south meanders (Fig. 11.49b) of the jet stream. These meanders are called **Rossby waves** or **planetary waves**. Typical wavelengths are 3000-4000 km. Given the circumference of a parallel at those latitudes, one typically finds 3 to 13 waves around the globe, with a normal **zonal wavenumber** of 7 to 8 waves. The equatorward region of any meander is called a **trough** (pronounced like “troff”) and is associated with low pressure or low geopotential height. The poleward portion of a meander is called a **ridge**, and has high pressure or height. The turning of winds around troughs and ridges are analogous to the turning around closed lows and highs, respectively. The trough center or **trough axis** is labelled with a dashed line, while the **ridge axis** is labelled with a zig-zag symbol (Fig. 11.49b).

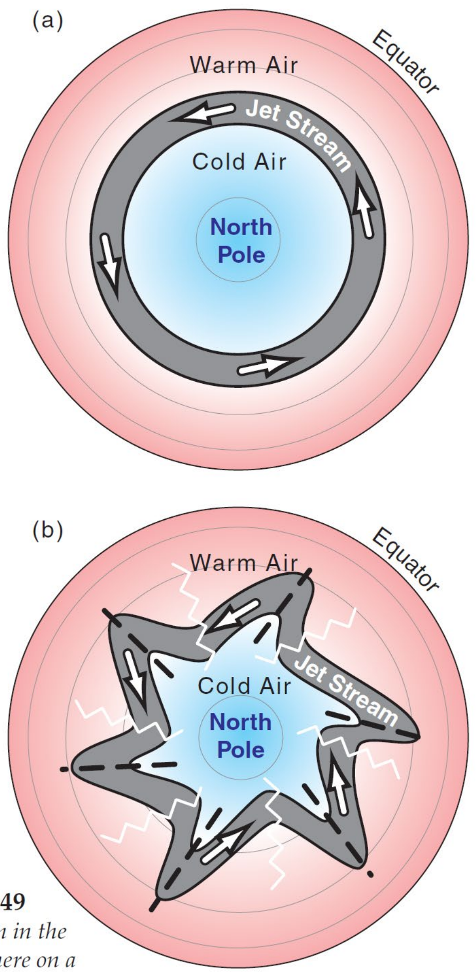
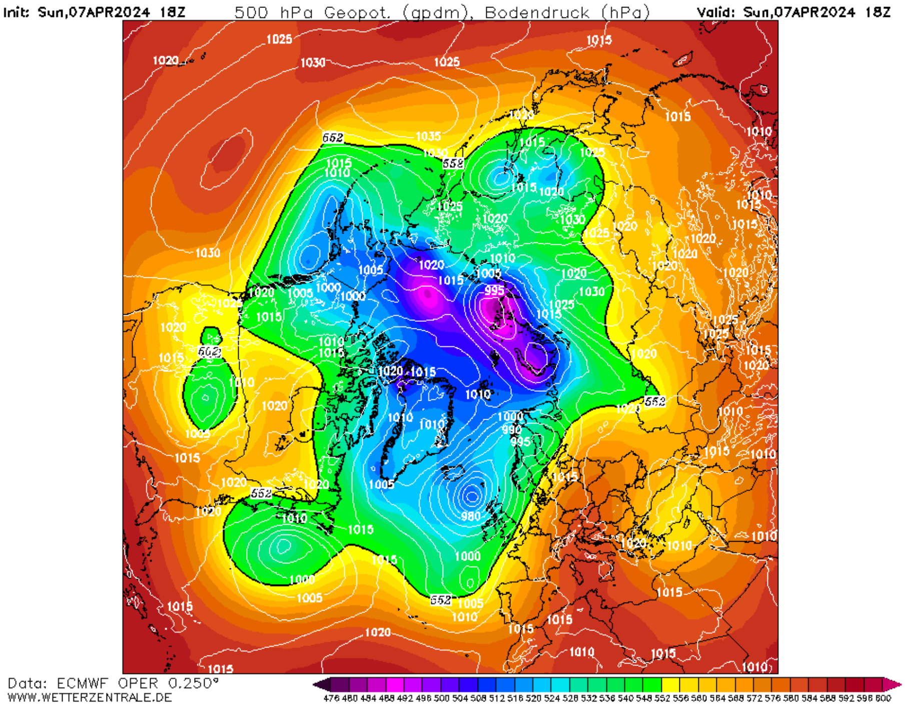
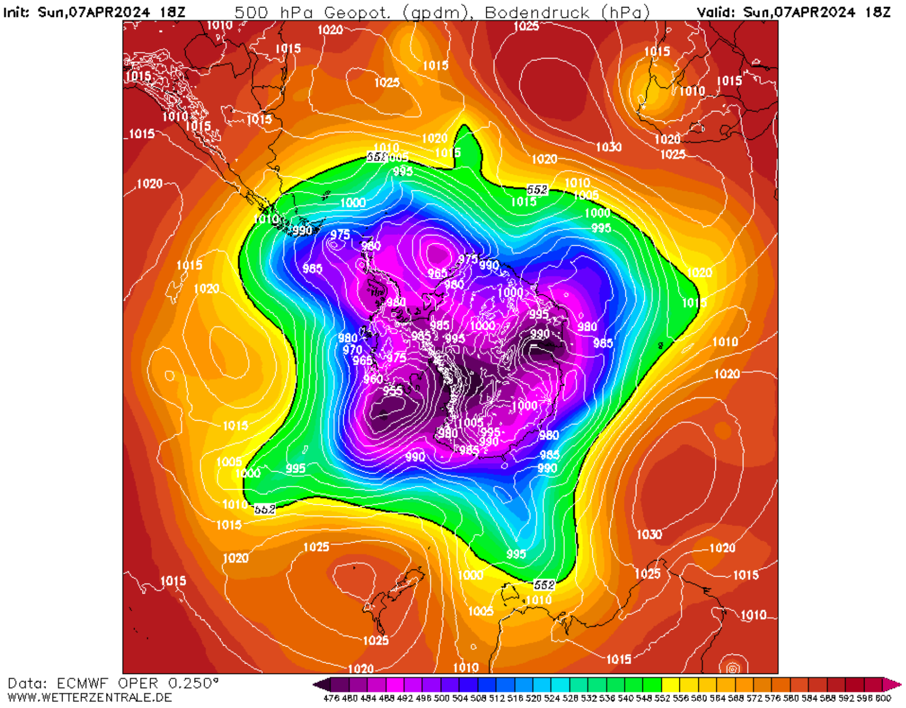


Figure 11.49
(a) Jet stream in the N. Hemisphere on a world with no instabilities. (b) Jet stream with barotropic or baroclinic instabilities, creating a meandering jet. Troughs are marked with a black dashed line, and ridges with a white zig-zag line.

Extratropical ridges & troughs (Rossby waves)

Today





North Hemisphere

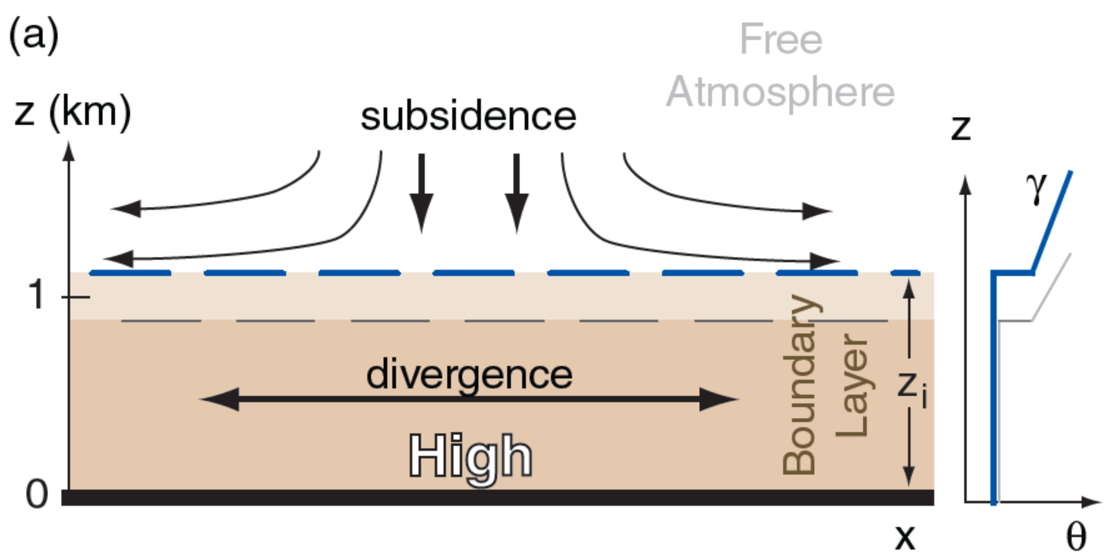
South Hemisphere

Anticyclones or highs

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.

Above a high center is often downward motion (**subsidence**) in the mid-troposphere, and horizontal spreading of air (**divergence**) near the surface

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



Anticyclones or highs

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. 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. They are seasonal (i.e., last for several months), and form due to the temperature contrast between land and ocean.

Anticyclones or highs

<u>Transient Rossby waves</u>: 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. They often exist for several days.

<u>Thunderstorms</u>: Downdrafts from thunderstorms create **meso-highs** roughly 10 to 20 km in diameter at the surface. These might exist for minutes to hours.

<u>Topography/Surface-Characteristics</u>: Mesohighs 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.

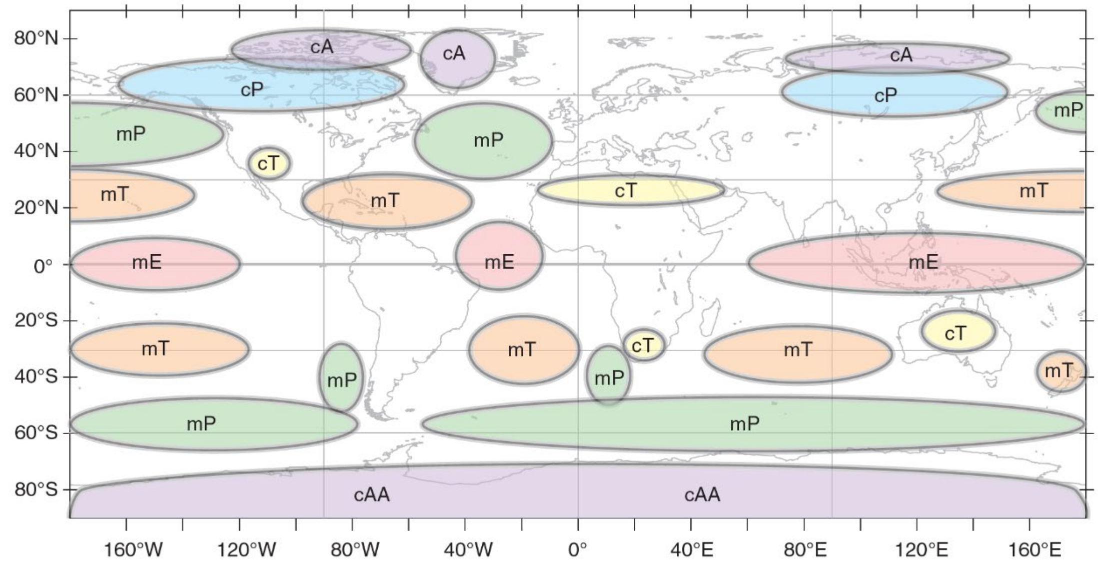
Airmasses

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

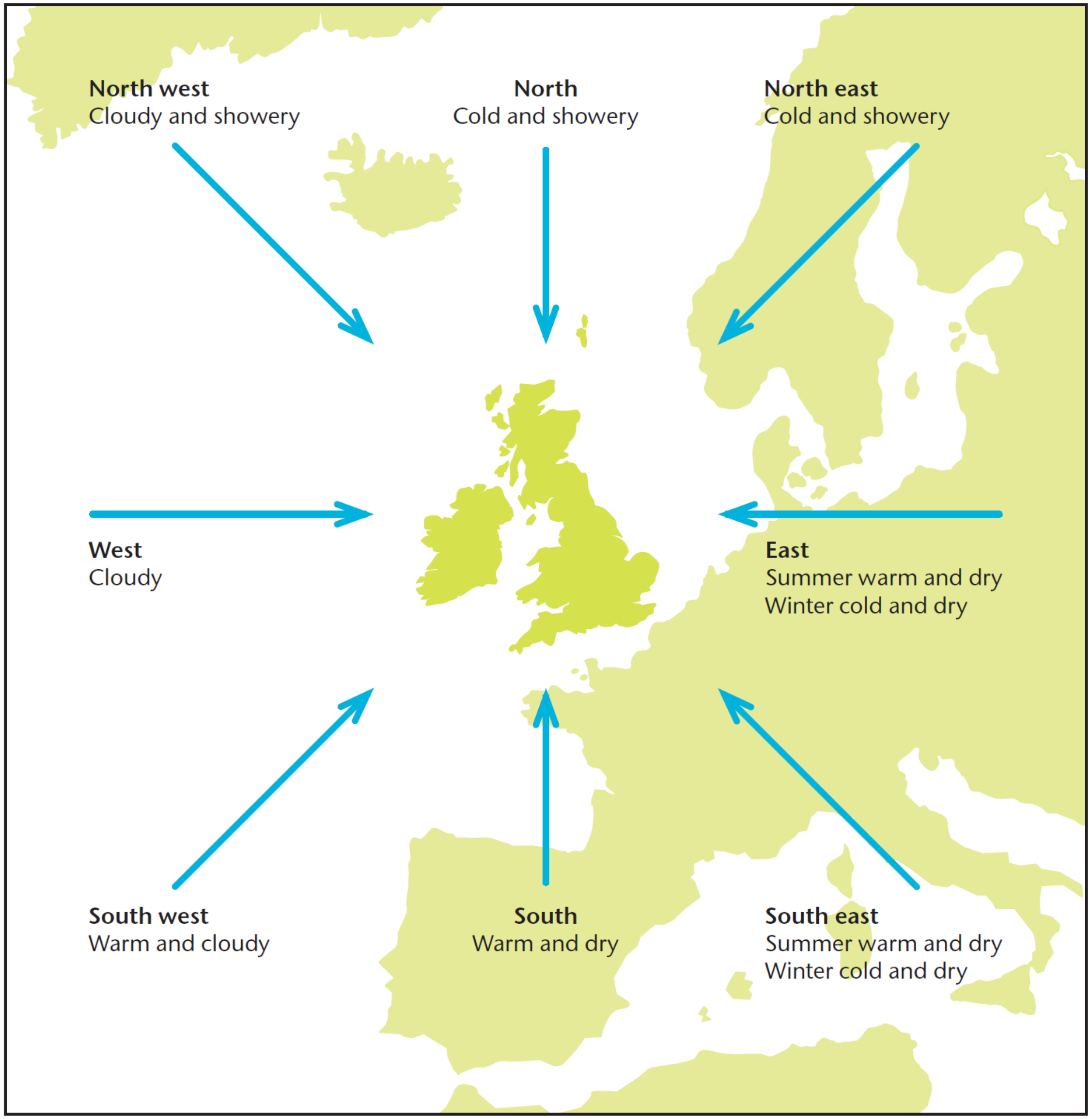
Abbr.	Name	Description
c	**continental**	Dry. Formed over land.
m	**maritime**	Humid. Formed over ocean.
A	Arctic	Very cold. Formed in the polar high.
E	Equatorial	Hot. Formed near equator.
M	Monsoon	Similar to tropical.
P	**Polar**	Cold. Formed in subpolar area.
S	Superior	A warm dry airmass having 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:

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

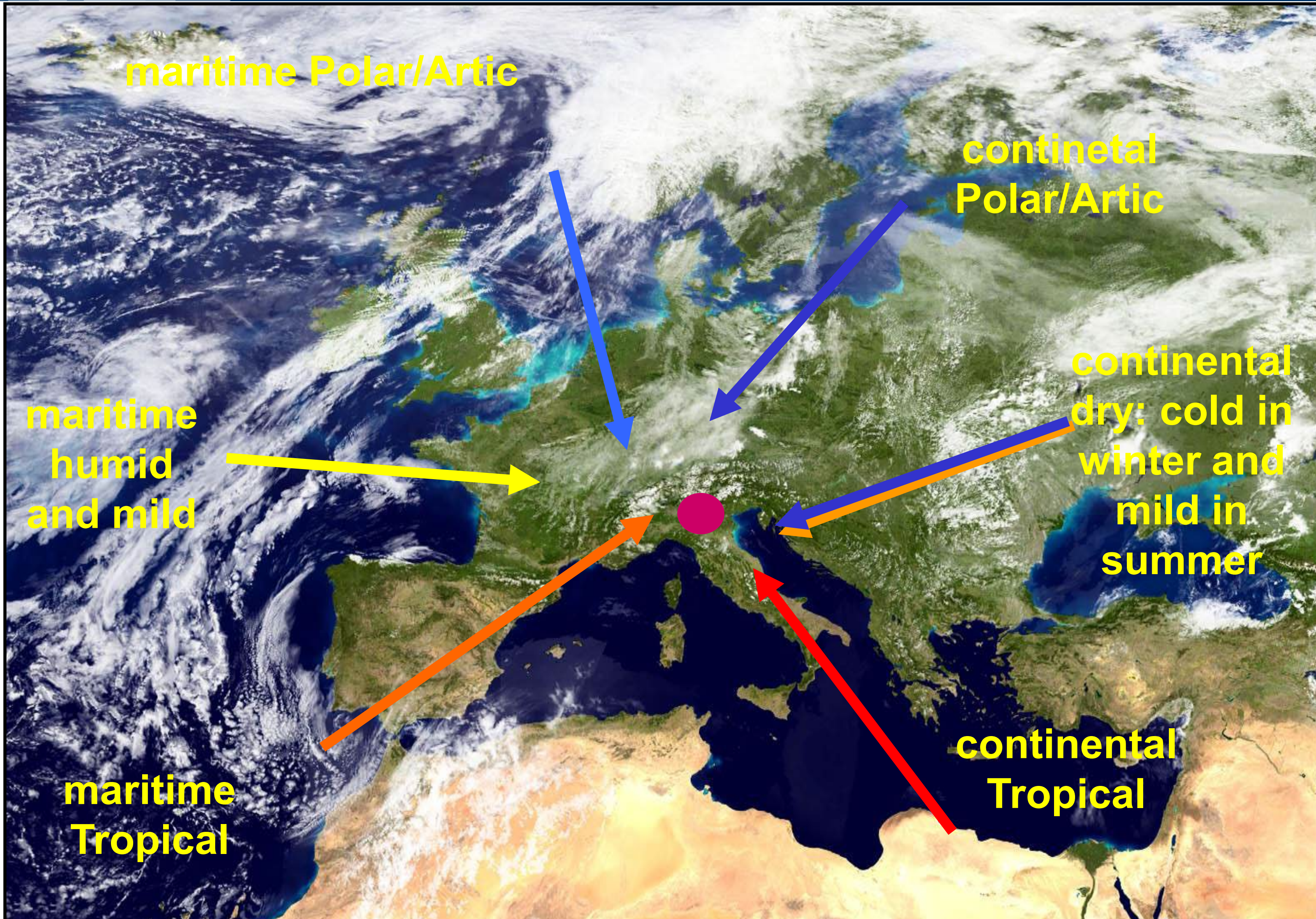


Airmasses



North west
Cloudy and showery
North
Cold and showery
North east
Cold and showery
West
Cloudy
East
Summer warm and dry
Winter cold and dry
South west
Warm and cloudy
South
Warm and dry
South east
Summer warm and dry
Winter cold and dry

Synoptic-scale climate drivers: air masses

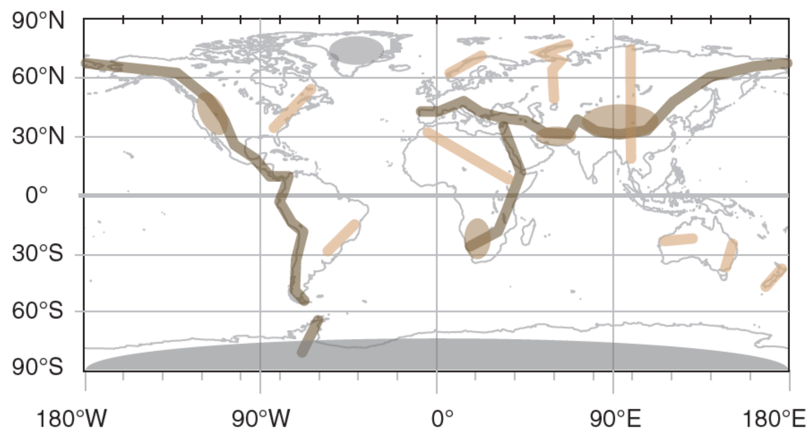


Airmasses

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 air masses move, two things can happen:
(1) As the air moves over surfaces with different characteristics, the airmass begins to change. This is called **air mass modification**

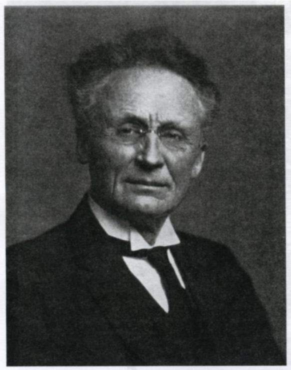
(2) An airmass can encounter another airmass. The boundary between these two air masses is called a **front**, and is a location of strong gradients of temperature, humidity, and other air mass characteristics.

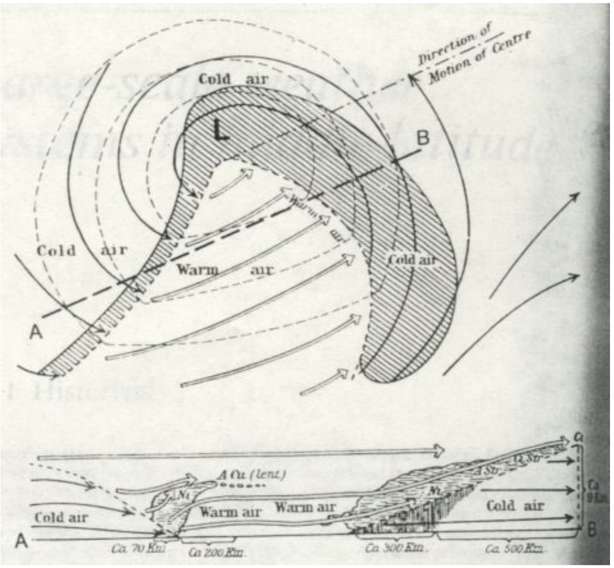
Tall mountain ranges can strongly **block** or **channel** the movement of air masses, because air masses occupy the bottom of the troposphere



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**.





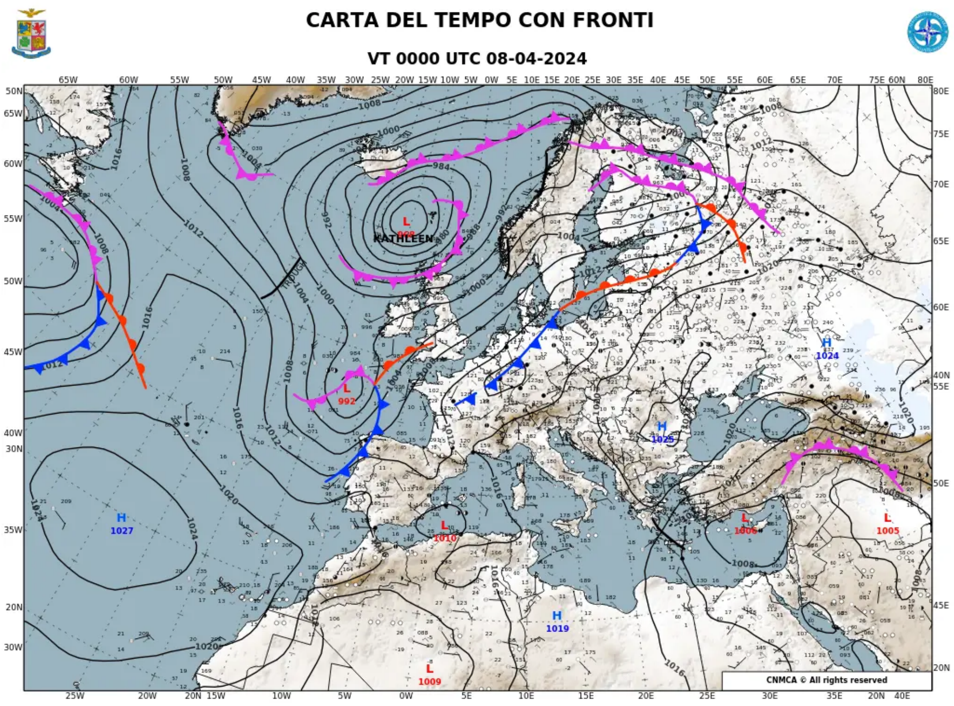
Surface fronts

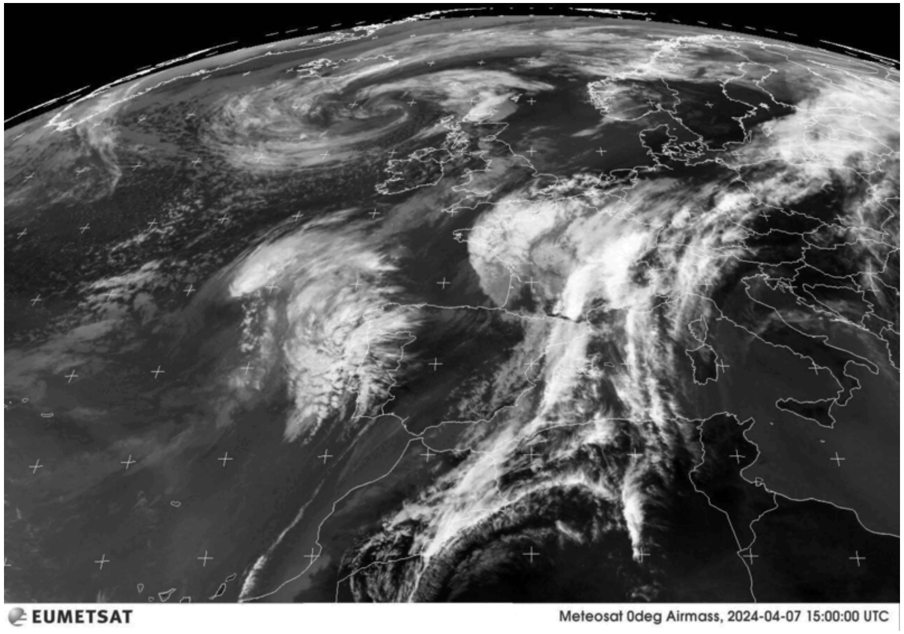
Surface fronts mark the boundaries between airmasses at the Earth's surface. They usually have the following attributes:

- **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.

Surface fronts



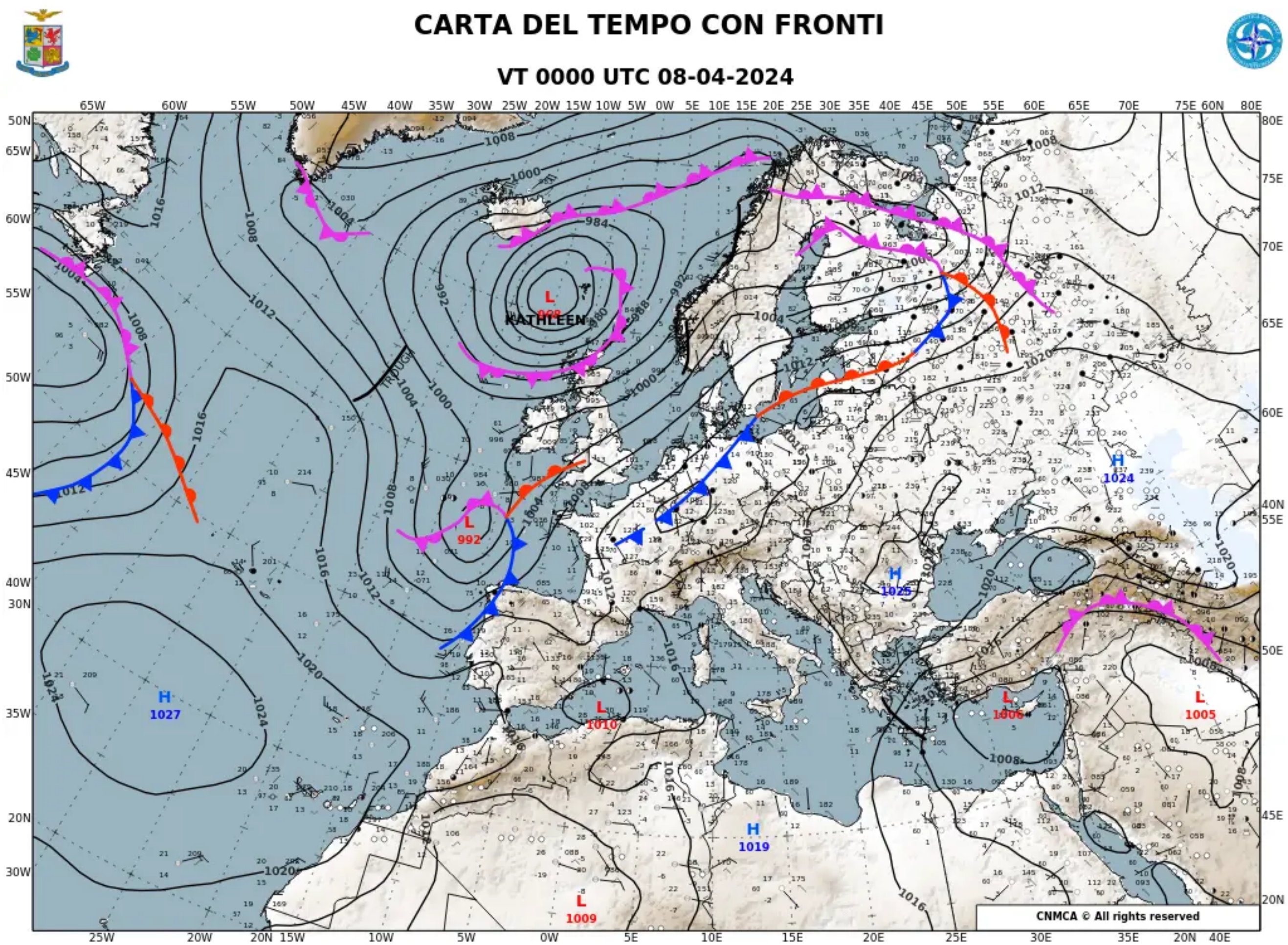


Surface fronts

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**.

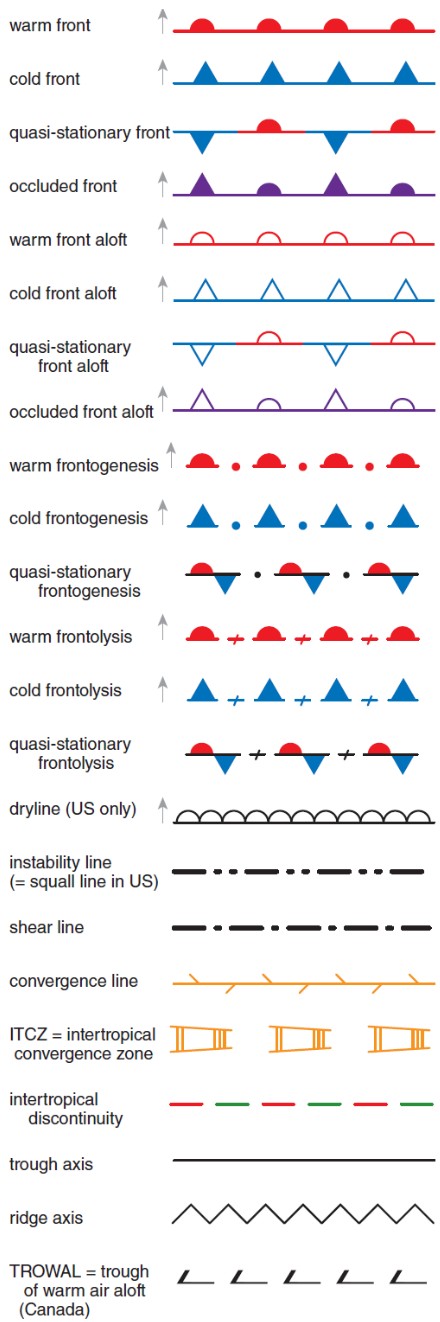
Fronts are always drawn on the warm side of the frontal zone. The frontal symbols are drawn on the side of the frontal line toward which the front is moving. For a stationary front, the symbols on both sides of the frontal line indicate what type of front it would be if it were to start moving in the direction the symbols point.

Surface pressure

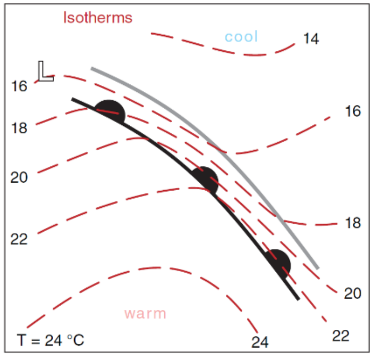


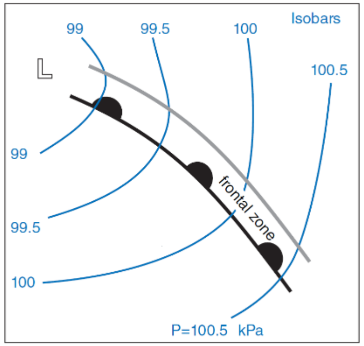
Surface fronts

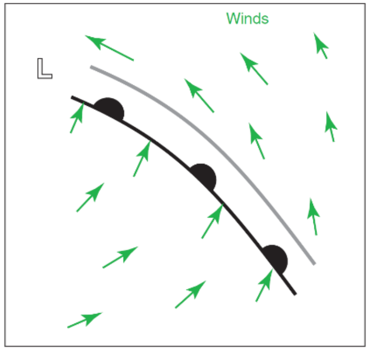
Glyphs for fronts, other airmass boundaries, and axes. The suffix "genesis" implies a forming or intensifying front, while "lysis" implies a weakening or dying front. A quasi-stationary front is a frontal boundary that doesn't move very much.

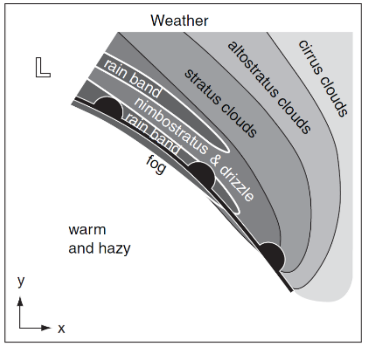


Warm fronts









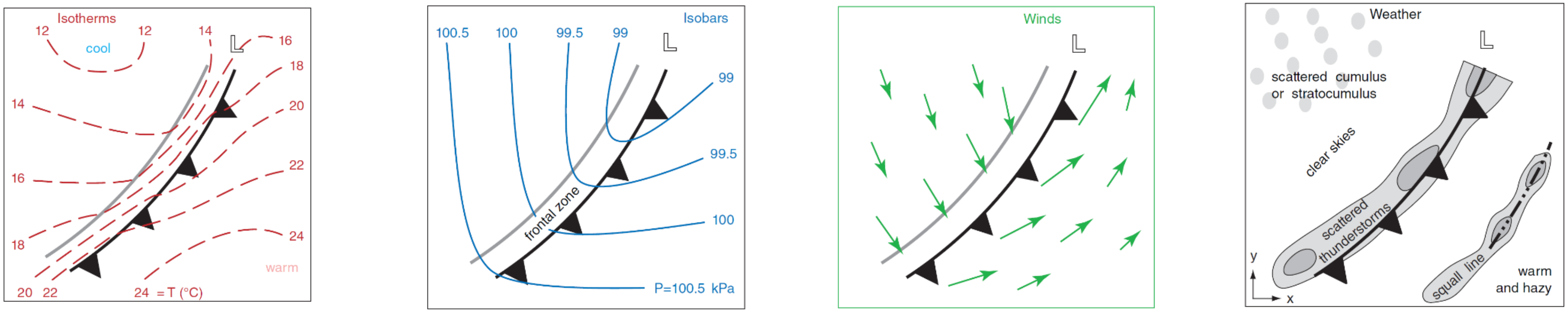
In central N. America, southeasterly winds ahead of the front bring in cool, humid air from the Atlantic Ocean, or bring in mild, humid air from the Gulf of Mexico.

An extensive deck of stratiform clouds (called a **cloud shield**) can occur hundreds of kilometres 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.

Cold fronts

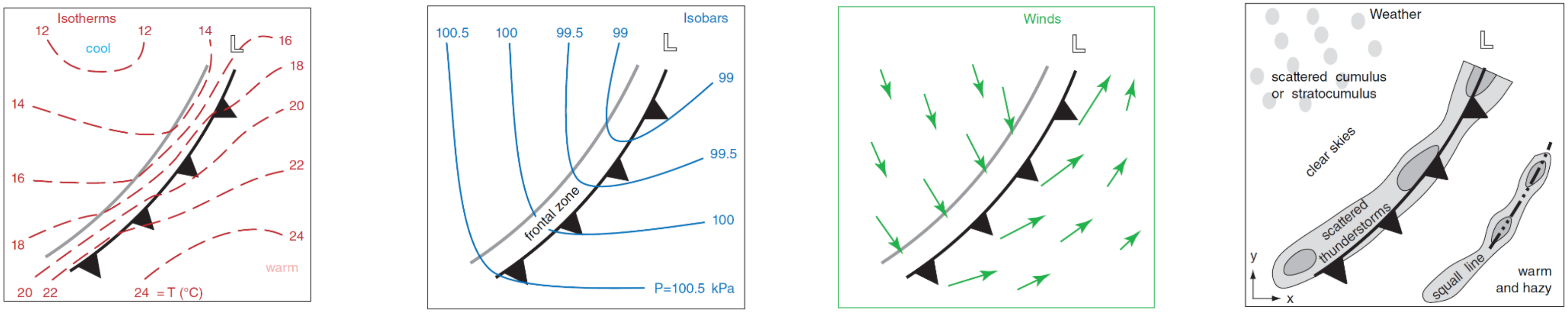


In central N. America, winds ahead of cold fronts typically have a southerly component, and can form strong low-level jets at night and possibly during day. Warm, humid, hazy air advects from the south. Sometimes a **squall line** of thunderstorms will form in advance of the front, in the warm air. These 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.

Cold fronts



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

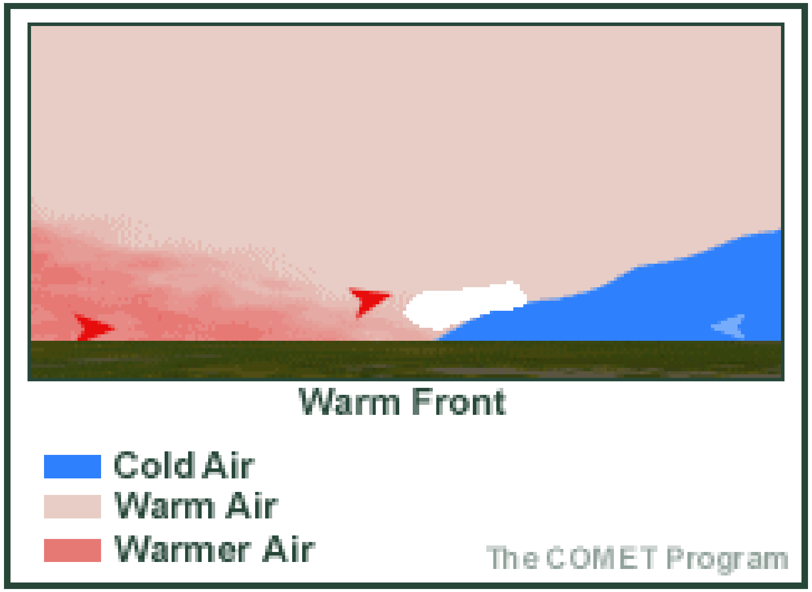
Sometimes over ocean surfaces the warm moist ocean leads to considerable post-frontal deep convection.

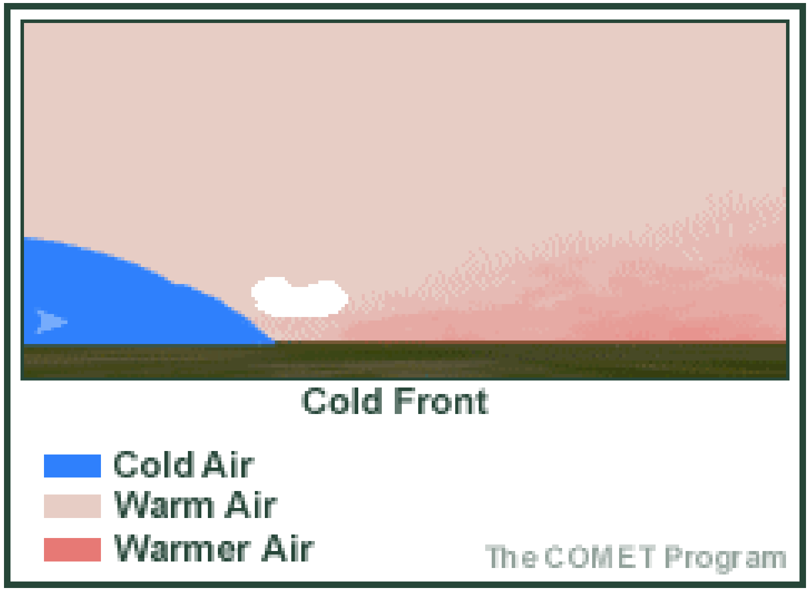
The idealized picture presented in figure can differ considerably in the mountains.

Surface fronts

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

The **cold front**, where the front moves toward the warm airmass (Fig. right). Retreating cold air defines the **warm front**, where the front moves toward the cold airmass (Fig. left).

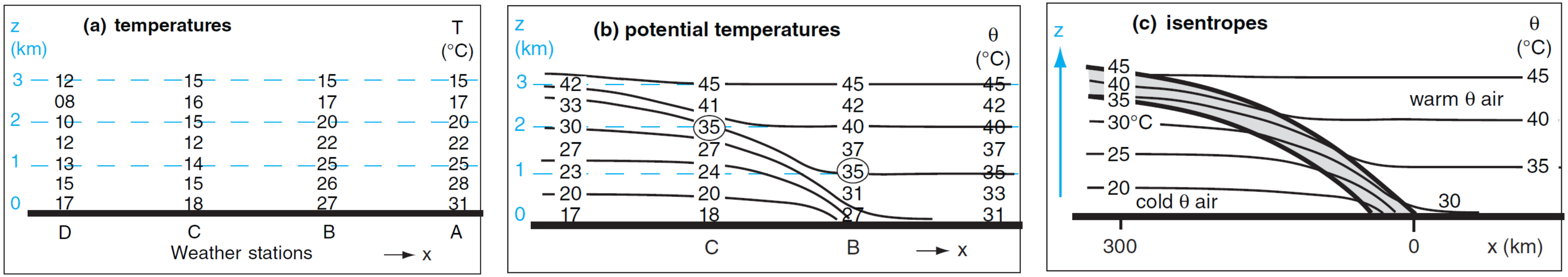




Fronts: vertical cross section

To locate fronts by their vertical cross section, first convert the temperatures into potential temperatures θ (Fig. b). Then, draw lines of equal potential temperature (**isentropes**). Fig. c shows isentropes re-drawn at 5°C intervals. Often isentropes are labelled in Kelvin.

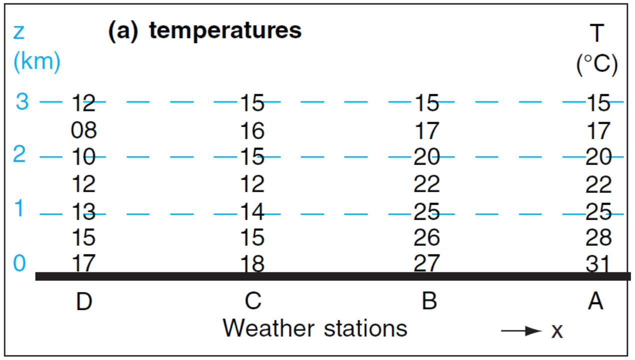
A frontal inversion is where the isentropes are packed closely together (shaded in Fig. c). This concept applies to both warm and cold fronts.

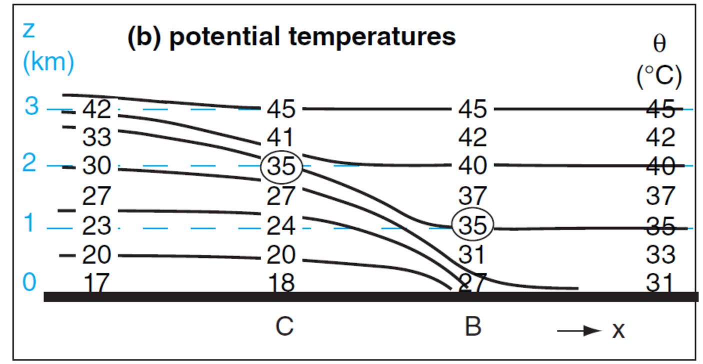


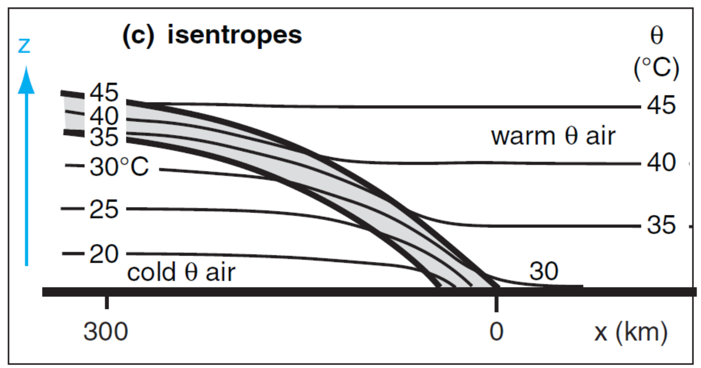
Air parcels that are forced to rise along isentropic surfaces can form clouds and precipitation, given sufficient moisture. Similarly, air blowing eastward would move downward along the sloping isentrope.

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.

Fronts: vertical cross section







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. b & c). The top of the cold air is called the **frontal inversion**

Surface fronts

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

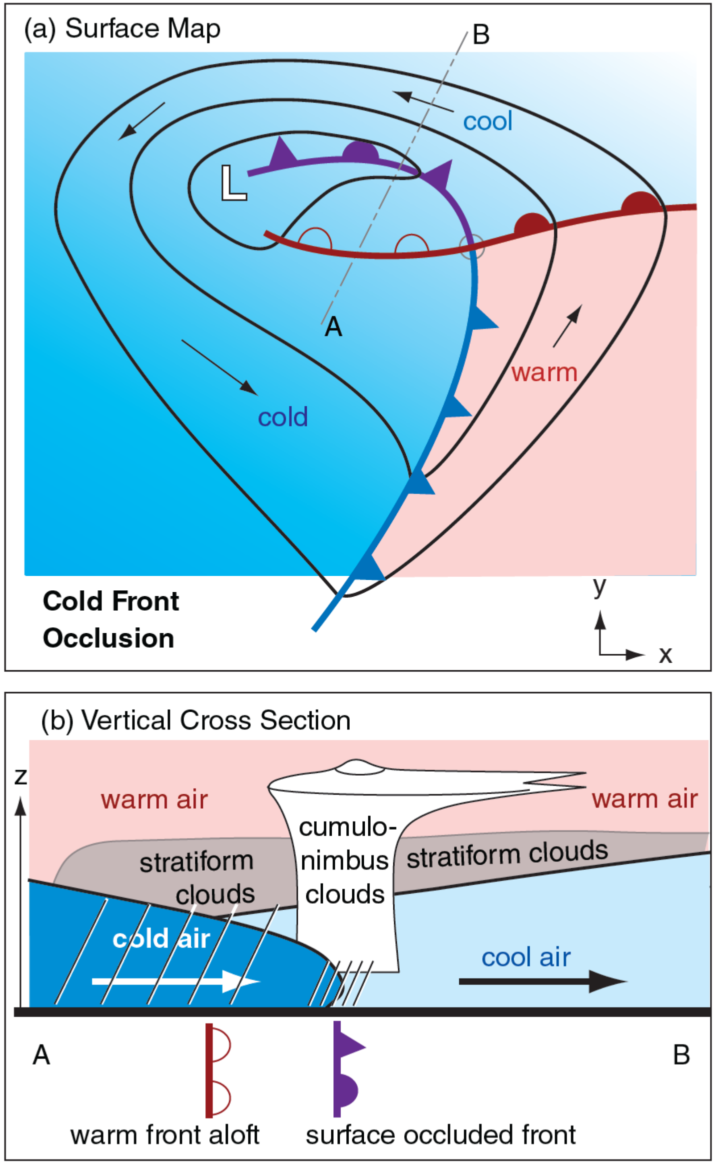
Fronts strengthen or weaken due to advection by the wind (kinematics), external diabatic heating (thermodynamics), and ageostrophic cross-frontal circulations (dynamics). One measure of frontal strength is the temperature change across it.

Surface fronts: occluded front

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.

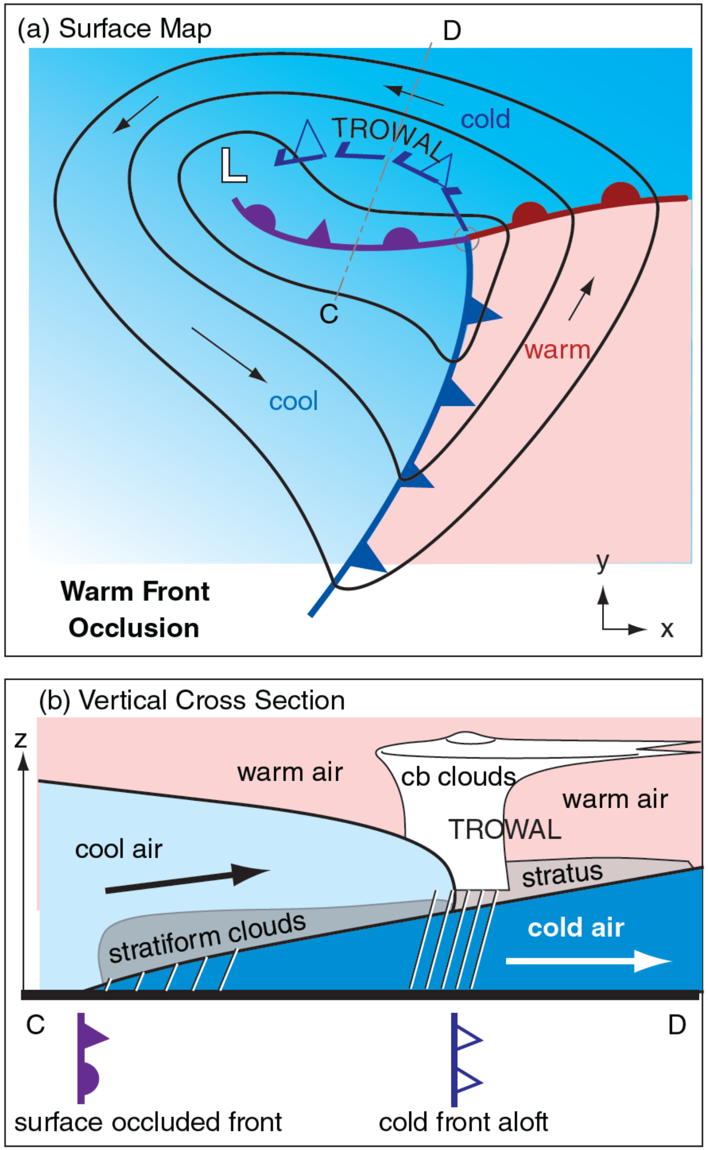
A **cold front occlusion**, where very cold air that is very statically stable catches up to, and under-rides, cooler air that is less statically stable. The warm air that was initially between these two cold air masses is forced aloft



Surface fronts: occluded front

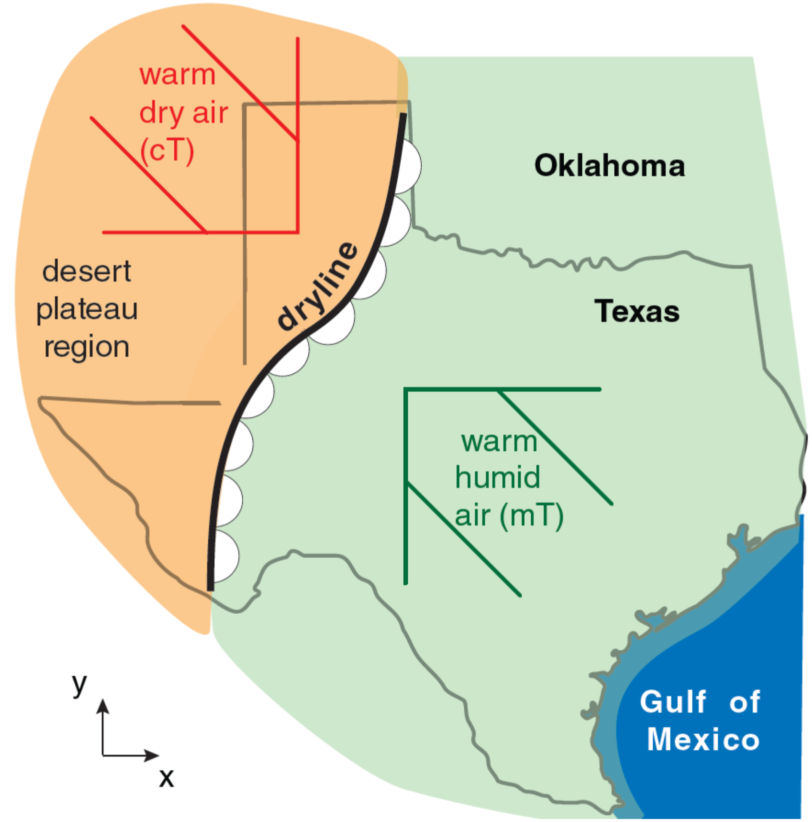
A **warm front occlusion**, where cool air that is less statically stable catches up to, and over-rides, colder air that is more statically stable, forcing aloft the warm air that was in between.

The wedge of warm air (Figure) pushed up between the cool and cold air masses is called a **TROWAL**, an acronym that means "**trough of warm air aloft**." This TROWAL, labeled in Fig. a, 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. b) — hence it is important for weather forecasting.



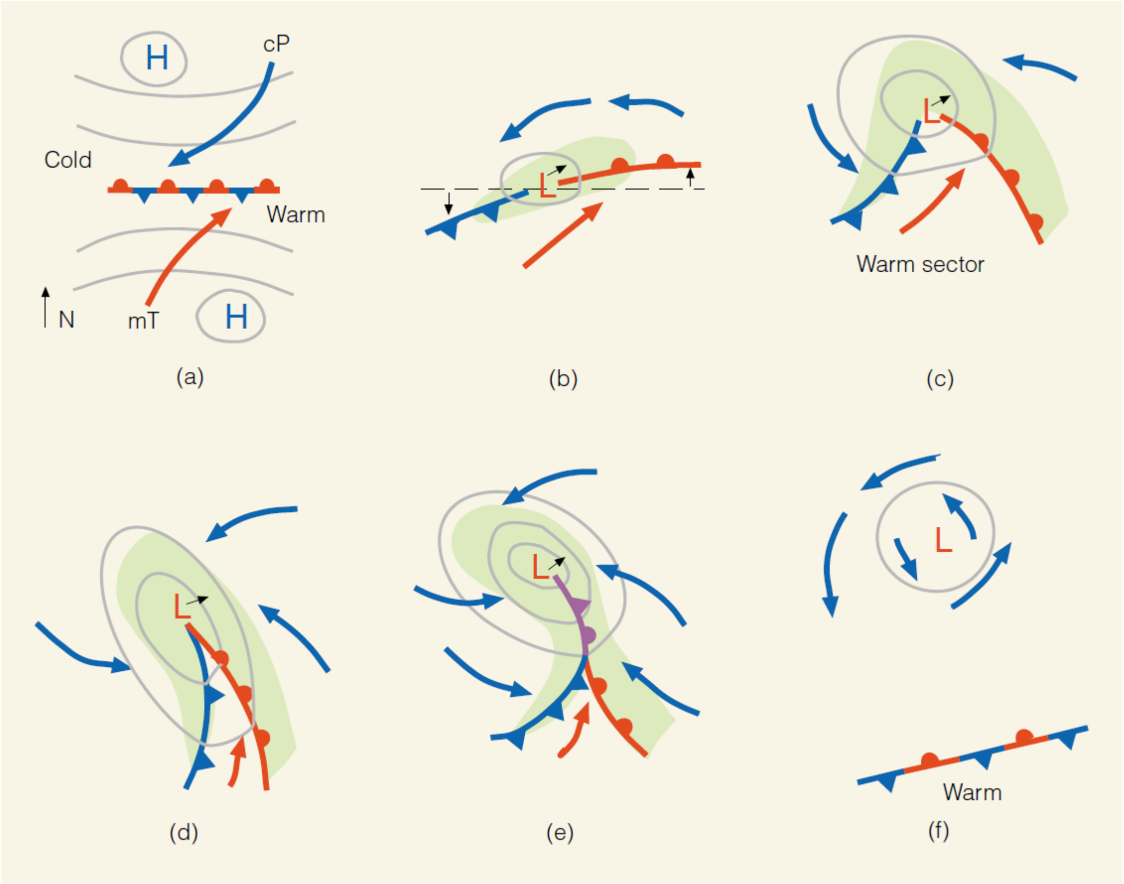
Dryline

There often exists a boundary between warm humid air to the east, and warm dry air to the west (Figure). 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.



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.

Surface fronts: life and death



The idealized life cycle of a wave cyclone (a through f) in the Northern Hemisphere based on the polar front theory. As the life cycle progresses, the system moves eastward in a dynamic fashion. The small arrow next to each L shows the direction of storm movement.

Extratropical cyclone

A synoptic-scale weather system with **low pressure** near the surface is called a **"cyclone"** (Figure). Horizontal winds turn **cyclonically** around it (clockwise/counterclockwise in the Southern/Northern Hemisphere). Near the surface these turning winds also spiral towards the low center. Ascending air in the cyclone can create clouds and precipitation. **Extratropical cyclones** (cyclones outside of the tropics) include transient **mid-latitude cyclones** and **polar cyclones**. Other names for extratropical cyclones are **lows** or **low-pressure centers** (see Table 13-1). Low-altitude convergence draws together airmasses to form fronts, along which the bad weather is often concentrated. These lows have a short life cycle (a few days to a week) as they are blown from west to east and poleward by the polar jet stream.

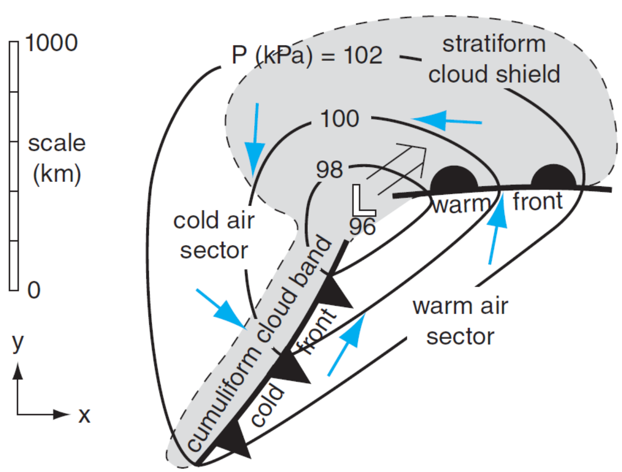


Table 13-1. Cyclone names. "Core" is storm center. *T* is relative temperature.

Common Name in N. Amer.	Formal Name	Other Common Names	T of the Core	Map Symbol
low	extra-tropical cyclone	mid-latitude cyclone	cold	L
		low-pressure center		
		storm system*		
		cyclone *(in N. America)*		
hurricane	tropical cyclone	typhoon *(in W. Pacific)*	warm	(hurricane symbol)
		cyclone *(in Australia)*		

(Often used by TV meteorologists.)*

Cyclogenesis & Cyclolysis

Cyclones are born and intensify (**cyclogenesis**) and later weaken and die (**cyclolysis**).

During cyclogenesis
(1) vorticity (horizontal winds turning around the low center)
(2) updrafts (vertical winds) increase
(3) surface pressure decreases.

In a nutshell, updrafts over a synoptic-scale region remove air from near the surface, causing the air pressure to decrease. The pressure gradient between this low-pressure center and the surroundings drives horizontal winds, which are forced to turn because of Coriolis force.
Frictional drag near the ground causes these winds to spiral in towards the low center, adding more air molecules horizontally to compensate for those being removed vertically. If the updraft weakens, the inward spiral of air molecules fills the low to make it less low (cyclolysis).

Cyclogenesis & Cyclolysis

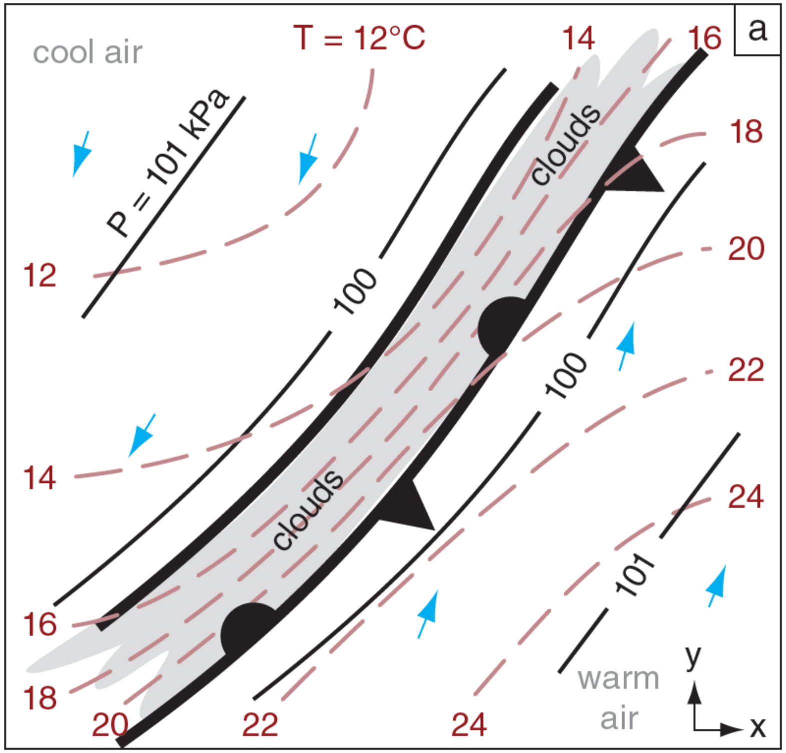
Cyclogenesis is enhanced at locations where one or more of the following conditions occur:

(1) east of mountain ranges, where terrain slopes downhill under the jet stream.
(2) east of deep troughs (and west of strong ridges) in the polar jet stream, where horizontal divergence of winds drives mid-tropospheric updrafts.
(3) at frontal zones or other baroclinic regions where horizontal temperature gradients are large.
(4) at locations that don't suppress vertical motions, such as where static stability is weak.
(5) where cold air moves over warm, wet surfaces such as the Gulf Stream, such that strong evaporation adds water vapor to the air and strong surface heating destabilizes the atmosphere.
(6) at locations further from the equator, where Coriolis force is greater.

Cyclone evolution

Although cyclones have their own synoptic-scale winds circulating around the low-pressure center, this whole system is blown toward the east by even larger-scale winds in the general circulation such as the jet stream.

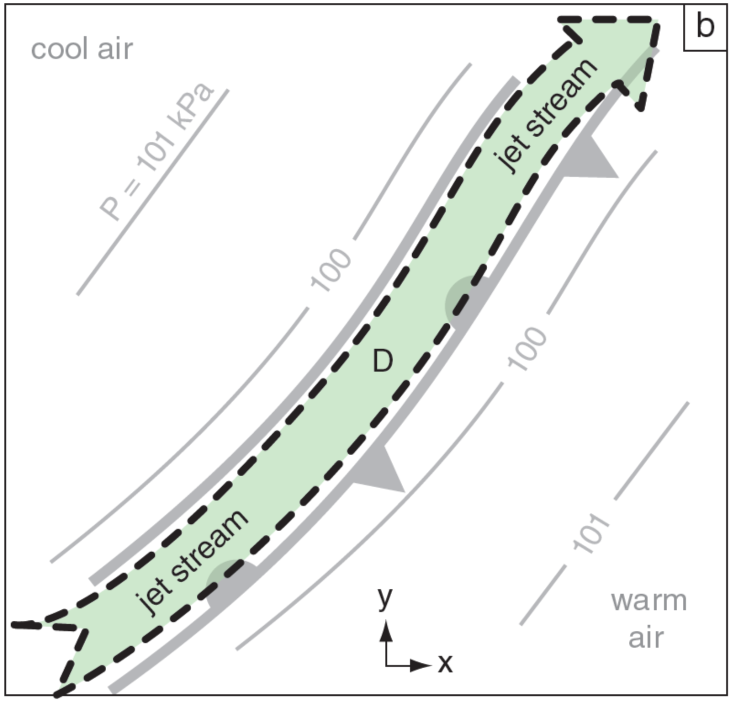
One condition that favors cyclogenesis is a **baroclinic** zone – a long, narrow region of large temperature change across a short horizontal distance near the surface



Cyclone evolution

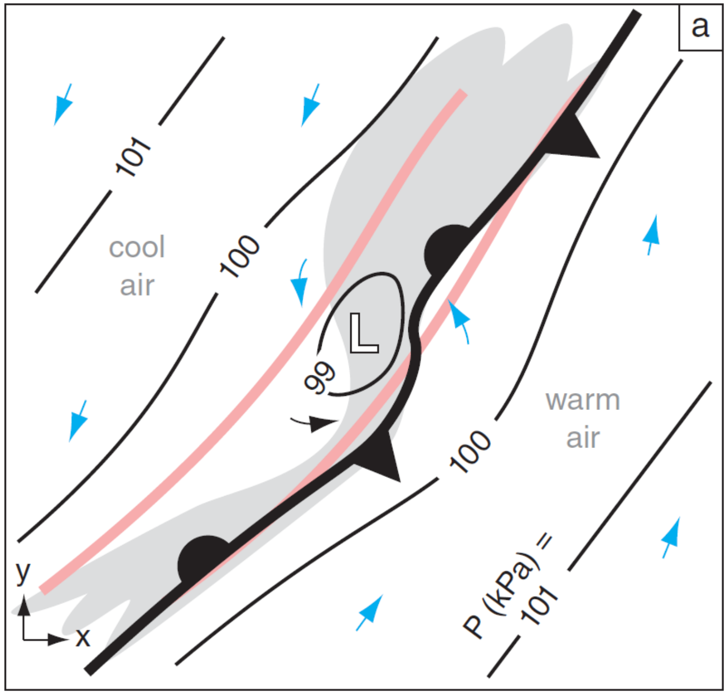
Above (near the tropopause) and parallel to this baroclinic zone is often a strong jet stream (Fig. b), driven by the thermal-wind effect.

If conditions are right, the jet stream can remove air molecules from a column of air above the front, at location “D” in Fig. b. This lowers the surface pressure under location “D”, causing **cyclogenesis** at the surface. Namely, under location “D” is where you would expect a surface low-pressure center to form.



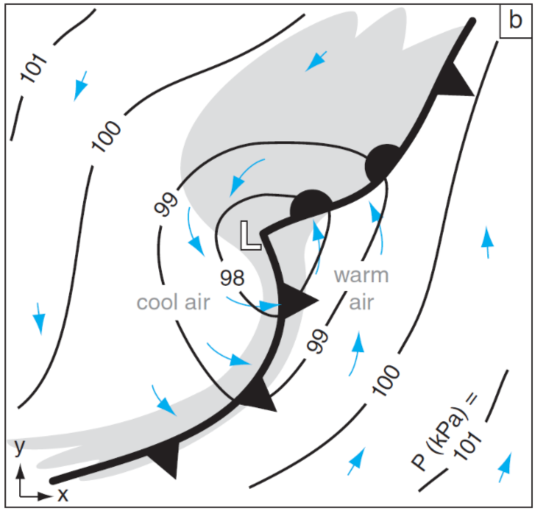
Cyclone evolution

The resulting pressure gradient around the surface low starts to generate lower-tropospheric winds that circulate around the low (Fig. a, again near the Earth's surface). This is the **spin-up** stage — so named because vorticity is increasing as the cyclone intensifies. The winds begin to advect the warm air poleward on the east side of the low and cold air equatorward on the west side, causing a kink in the former stationary front near the low center. The kinked front is wave shaped, and is called a **frontal wave**. Parts of the old front advance as a warm front, and other parts advance as a cold front. Also, these winds begin to force some of the warmer air up over the colder air, thereby generating more clouds.

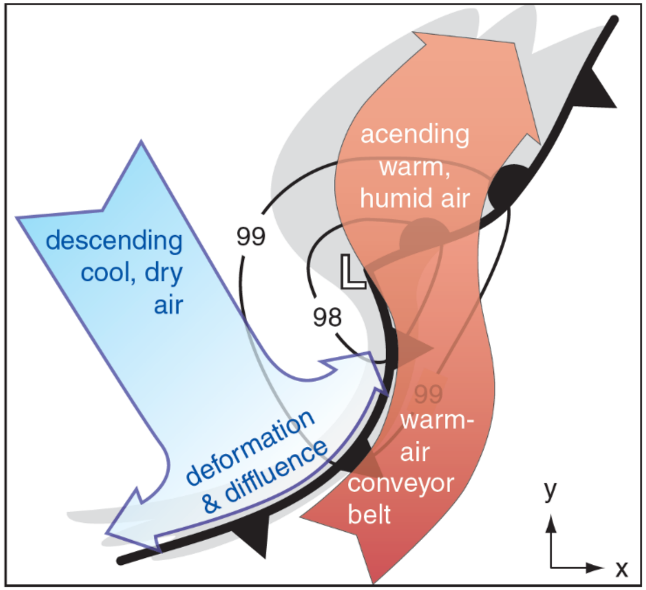


Cyclone evolution

If jet-stream conditions continue to be favorable, then the low continues to intensify and mature (Fig. b). As this cyclogenesis continues, the central pressure drops (namely, the cyclone **deepens**), and winds and clouds increase as a **vortex** around the **low center**. Precipitation begins if sufficient moisture is present in the regions where air is rising.

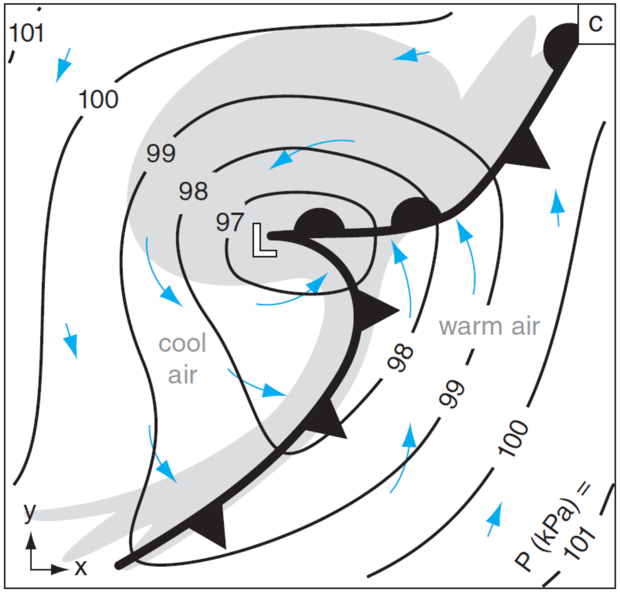


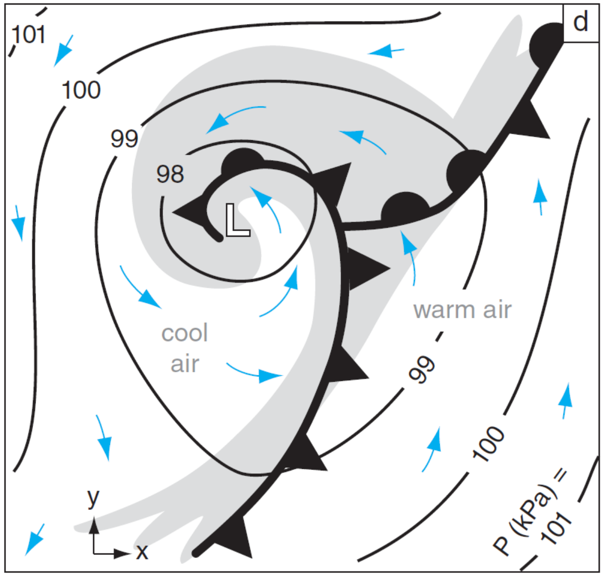
Sometimes air in the **warm-air conveyor belt** is moving so fast that it is called a low-altitude **pre-frontal jet**. When this humid stream of air is forced to rise over the cooler air at the warm front (or over a mountain) it can dump heavy precipitation and cause flooding.



Cyclone evolution

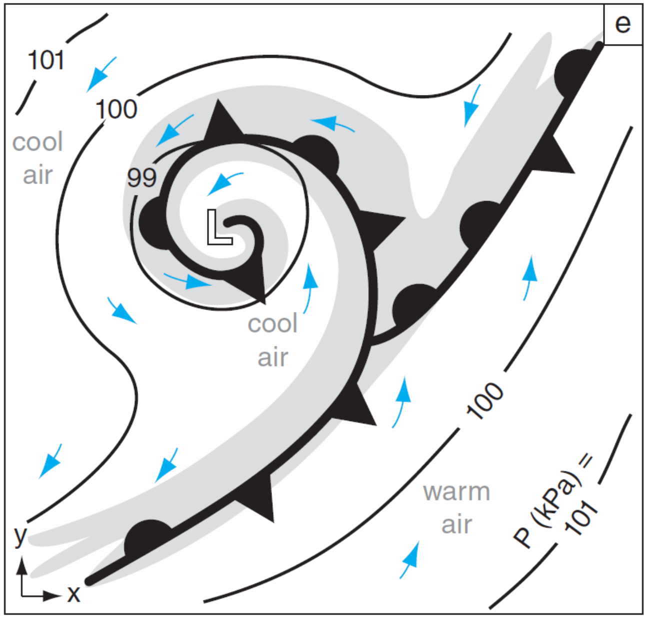
At the peak of cyclone intensity (lowest central pressure and strongest surrounding winds) the cold front often catches up to the warm front near the low center (Fig. c). As more of the cold front overtakes the warm front, an occluded front forms near the low center (Fig. d). The cool air is often drier, and is visible in satellite images as a **dry tongue** of relatively cloud-free air that begins to wrap around the low. This marks the beginning of the **cyclolysis** stage. During this stage, the low is said to **occlude** as the occluded front wraps around the low center.





As the cyclone occludes further, the low center becomes surrounded by cool air (Fig. e). Clouds during this stage spiral around the center of the low — a signature that is easily seen in satellite images.

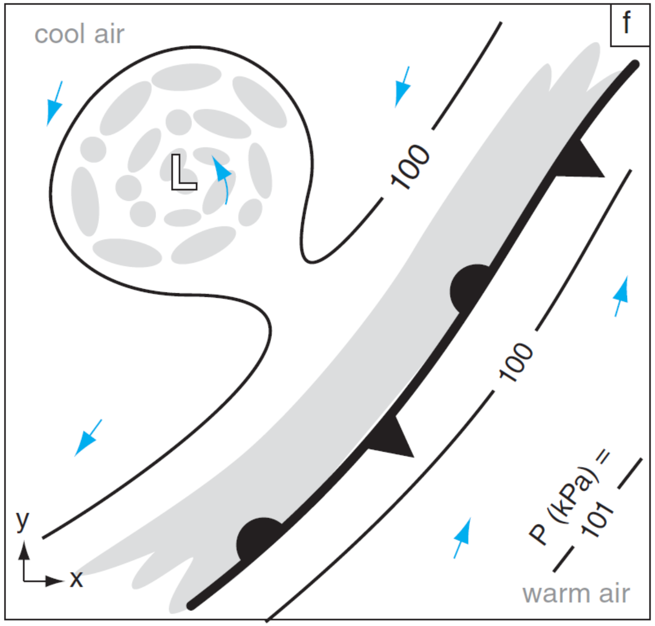
But the jet stream, still driven by the thermal wind effect, moves east of the low center to remain over the strongest baroclinic zone (over the warm and cold fronts, which are becoming more stationary).



Cyclone evolution

Without support from the jet stream to continue removing air molecules from the low center, the low begins to **fill** with air due to convergence of air in the boundary layer. The central pressure starts to rise and the winds slow as the vorticity **spins down**.

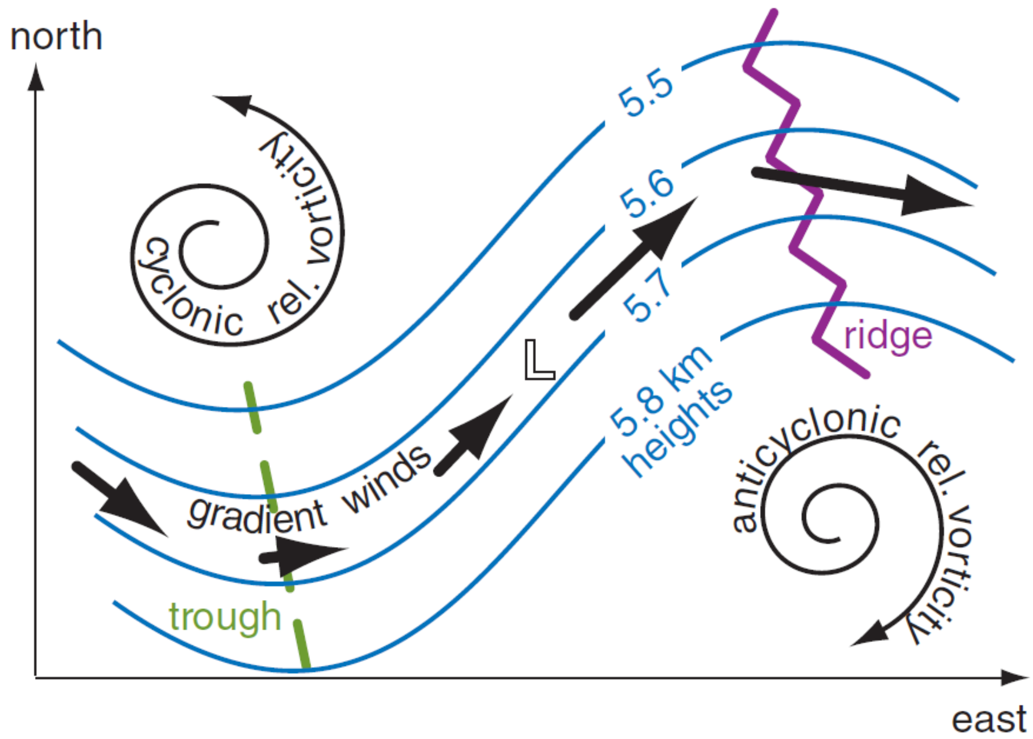
As cyclolysis continues, the low center often continues to slowly move further poleward away from the baroclinic zone (Fig. f). The central pressure continues to rise and winds weaken. The tightly wound spiral of clouds begins to dissipate into scattered clouds, and precipitation diminishes.



While they exist, they are driven by the baroclinicity in the air (through the action of the jet stream).
But their circulation helps to reduce the baroclinicity by moving cold air equatorward, warm air poleward, and mixing the two airmasses together. As described by **Le Chatelier's Principle** (namely, an imbalanced system reacts in a way to partially counteract the imbalance), the cyclone forms as a response to the baroclinic instability, and its existence partially undoes this instability. Namely, cyclones help the global circulation to redistribute heat between equator and poles.

Cyclones are often strengthened in regions under the jet stream just east of troughs. In such regions, the jet stream steers the low center toward the east and poleward. Hence, cyclone tracks are often toward the northeast in the N. Hemisphere, and toward the southeast in the S. Hemisphere.

Cyclones in the Northern Hemisphere typically evolve during a 2 to 7 day period, with most lasting 3 - 5 days. They travel at typical speeds of 12 to 15 m s^{-1} (43 to 54 km h^{-1}), which means they can move about 5000 km during their life.

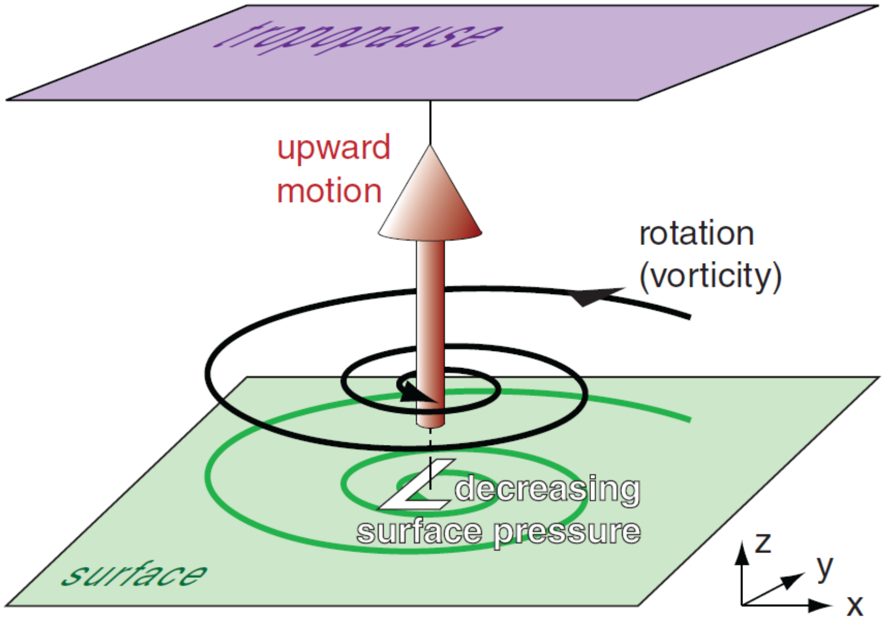


An idealized 50 kPa chart with equally-spaced height contours, as introduced by J. Bjerknes in 1937. The location of the surface low L is indicated.

Cyclogenesis is associated with:

(a) upward motion,
(b) decreasing surface pressure
(c) increasing vorticity (i.e., spin-up).

You can gain insight into cyclogenesis by studying all three characteristics, even though they are intimately related.



The processes that cause cyclogenesis (**spin up**; positive-vorticity increase) and cyclolysis (**spin down**; positive-vorticity decrease)

Omega equation

The definition of vertical motion is $W = \Delta z/\Delta t$, for altitude z and time t. Because each altitude has an associated pressure, define a new type of vertical velocity in terms of pressure. This is called **omega** (ω):

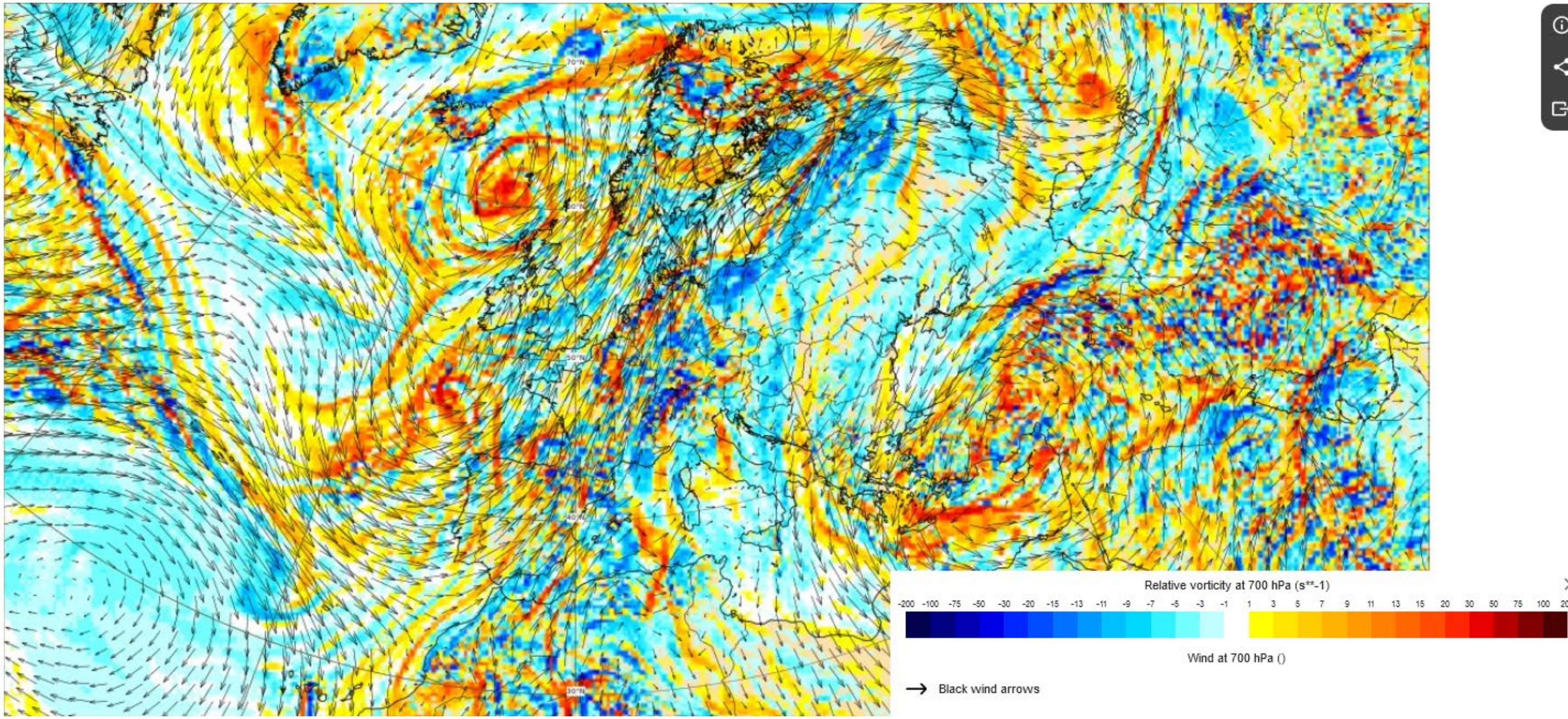
$$\omega = \frac{\Delta P}{\Delta t}$$

Omega has units of Pa s^{-1}.
You can use the hydrostatic equation to relate W and ω:

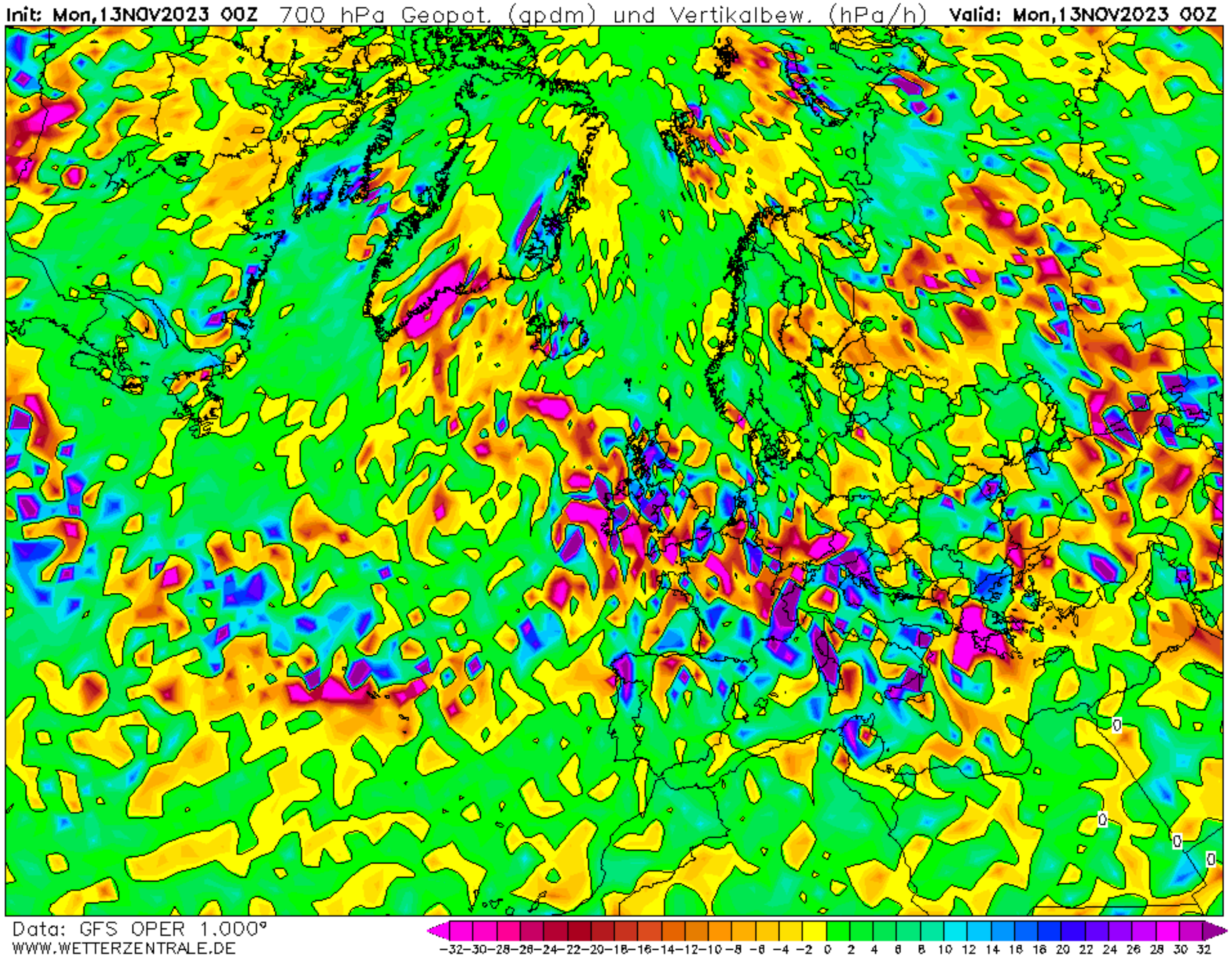
$$\omega = -\rho \cdot |g| \cdot W$$

for gravitational acceleration magnitude $|g| = 9.8$ m · s^{-2} and air density ρ. The negative sign in eq. implies that updrafts (positive W) are associated with negative ω. As an example, if your weather map shows $\omega = -0.68$ Pa s^{-1} on the 50 kPa surface, then the equation above can be rearranged to give $W = 0.1$ m s^{-1}, where a mid-tropospheric density of $\rho \approx 0.69$ kg · m^{-3} was used.
Use either W or ω to represent vertical motion. Numerical weather forecasts usually output the vertical velocity as ω.

700 hPa Geopotential height and vorticity



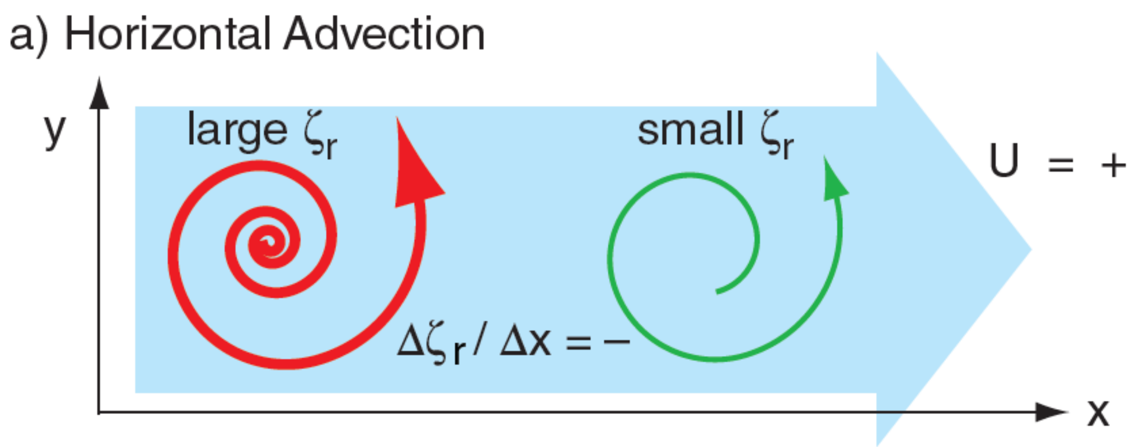
700 hPa Geopotential height and vertical motion



Positive vorticity tendency indicates cyclogenesis.

Vorticity Advection: If the wind blows air of greater vorticity into your region of interest, then this is called **positive vorticity advection** (**PVA**).

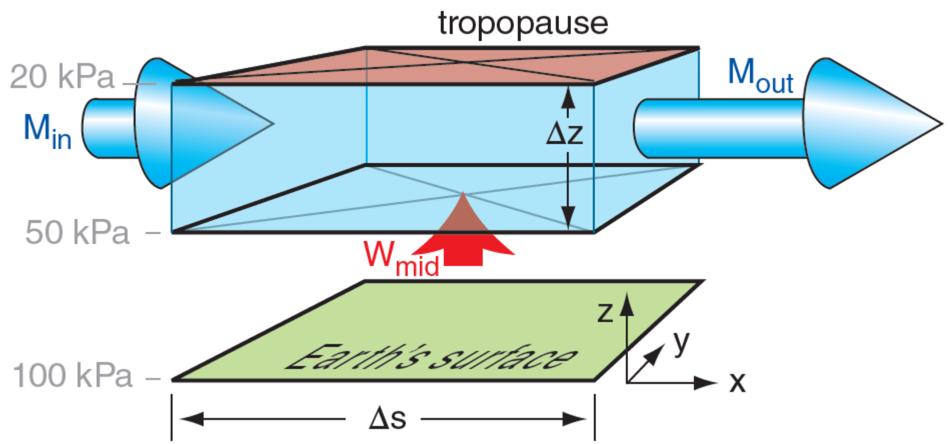
Negative vorticity advection (**NVA**) is when lower-vorticity air is blown into a region. These advections can be caused by vertical winds and horizontal winds



The importance of jet stream

Near the tropopause, horizontal divergence of jet-stream winds can force mid tropospheric ascent in order to conserve air mass as given by the continuity equation.

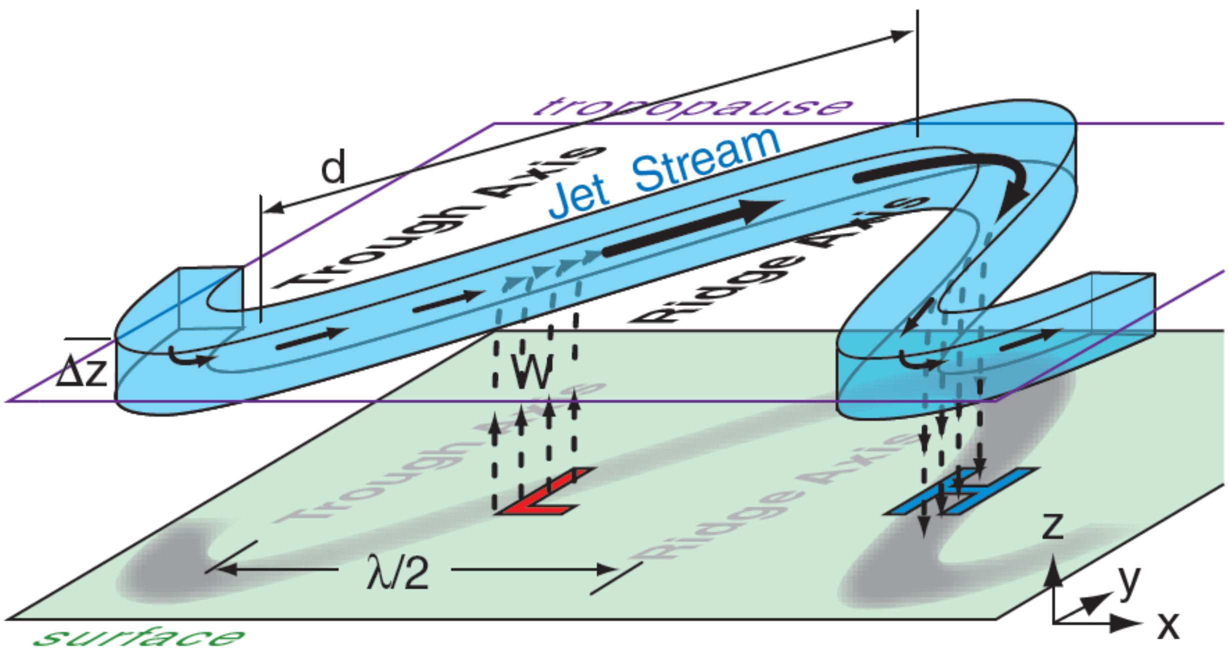
Horizontal divergence ($D = \Delta U/\Delta x + \Delta V/\Delta y$) is where more air leaves a volume than enters, horizontally.



Net vertical inflow can compensate for net horizontal outflow

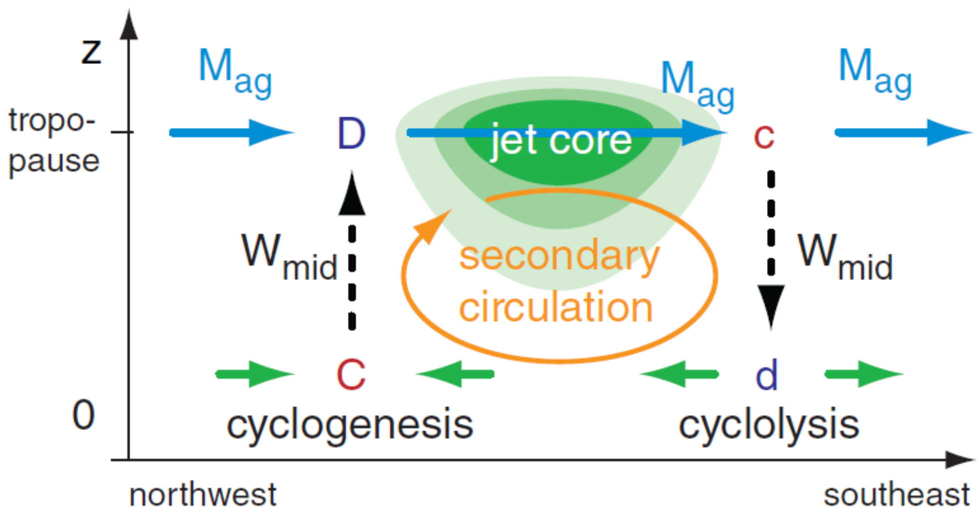
The importance of jet stream

Consider a hypothetical box of air at the jet stream level between the trough and ridge. Horizontal wind speed entering the box is slow around the trough, while exiting winds are fast around the ridge. To maintain mass continuity, this horizontal divergence induces ascent into the bottom of the hypothetical box. This ascent is removing air molecules below the hypothetical box, creating a region of low surface pressure. Hence, **surface lows** (extratropical cyclones) **form east of jet-stream troughs**.



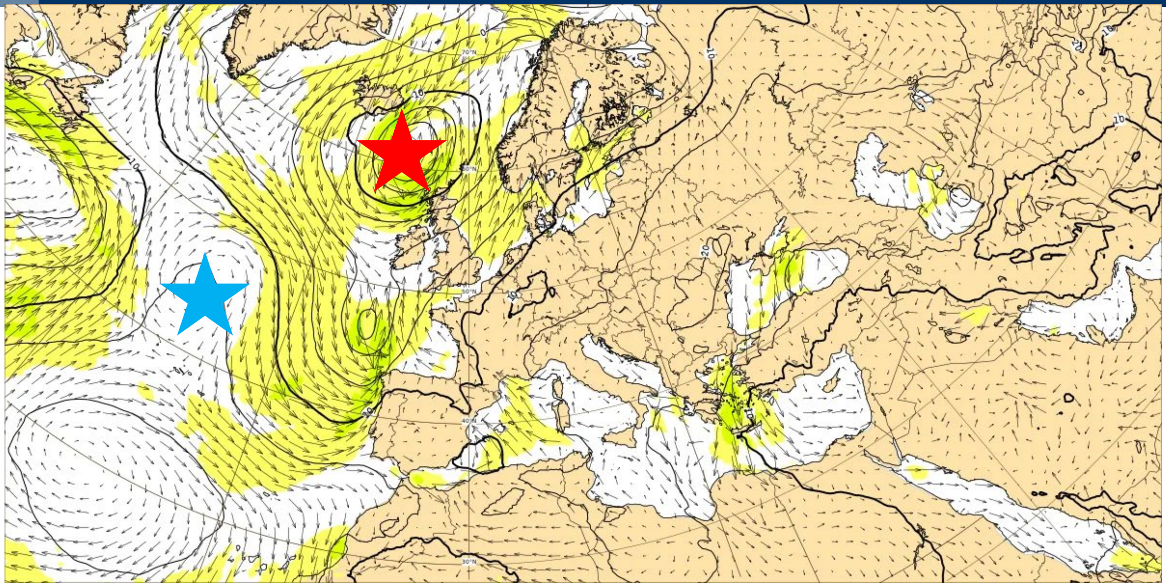
The importance of jet stream

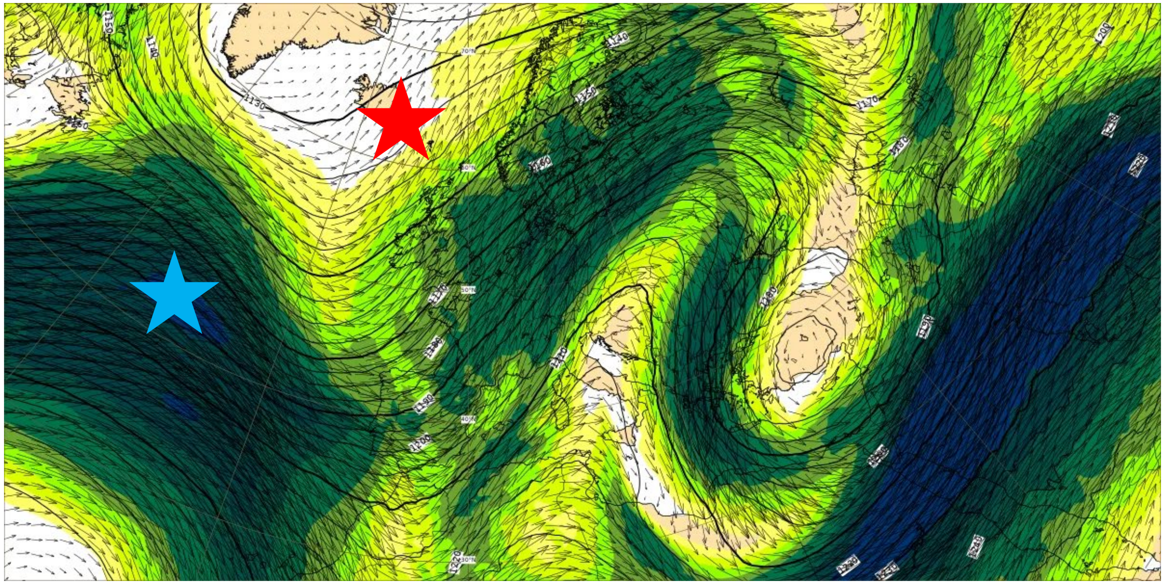
The jet stream does not maintain constant speed in the **jet core** (center region with maximum speeds). Instead, it accelerates and decelerates as it blows around the world in response to changes in horizontal pressure gradient and direction. The fast wind regions in the jet core are called **jet streaks**.

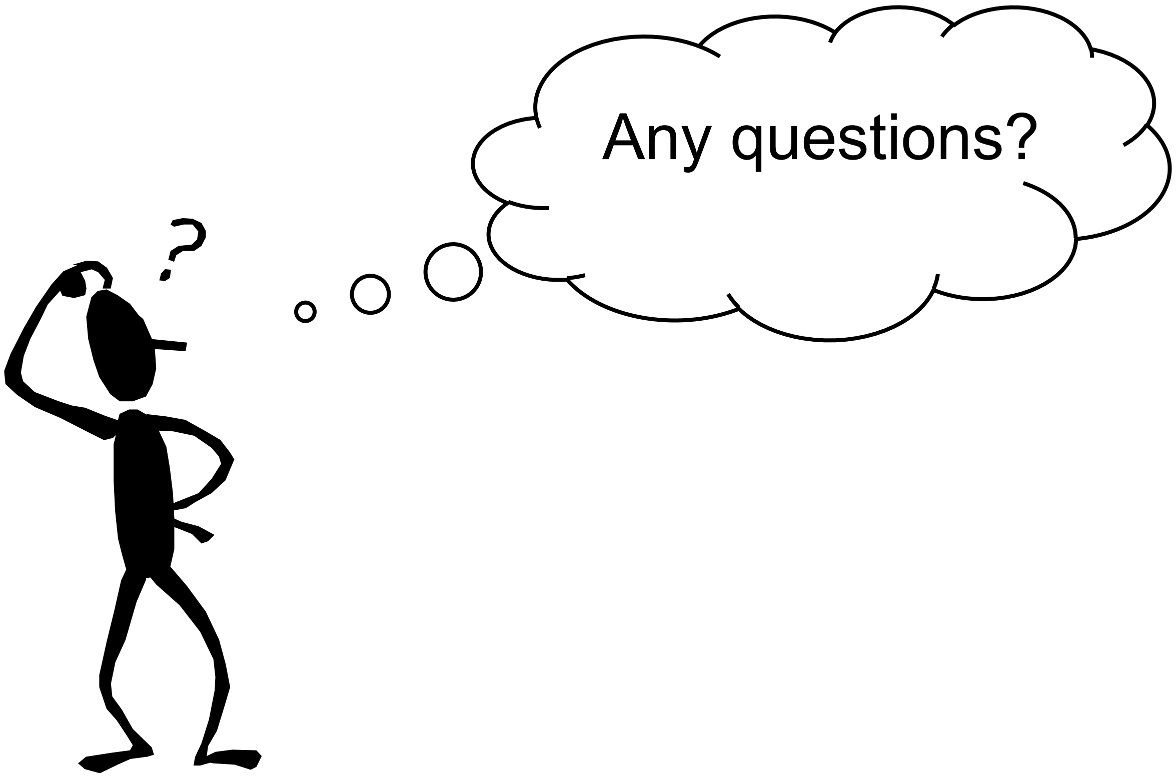


Vertical slice through the atmosphere at the jet-streak exit region, perpendicular to the average jet direction. Viewed from the west southwest, the green shading indicates isotachs of the jet core into the page. Divergence (D) in the left exit region creates ascent (W, dotted lines) to conserve air mass, which in turn removes air from near the surface. This causes the surface pressure to drop, favoring cyclogenesis. The opposite happens under the right exit region, where cyclolysis is favored.

The importance of jet stream






Any questions?