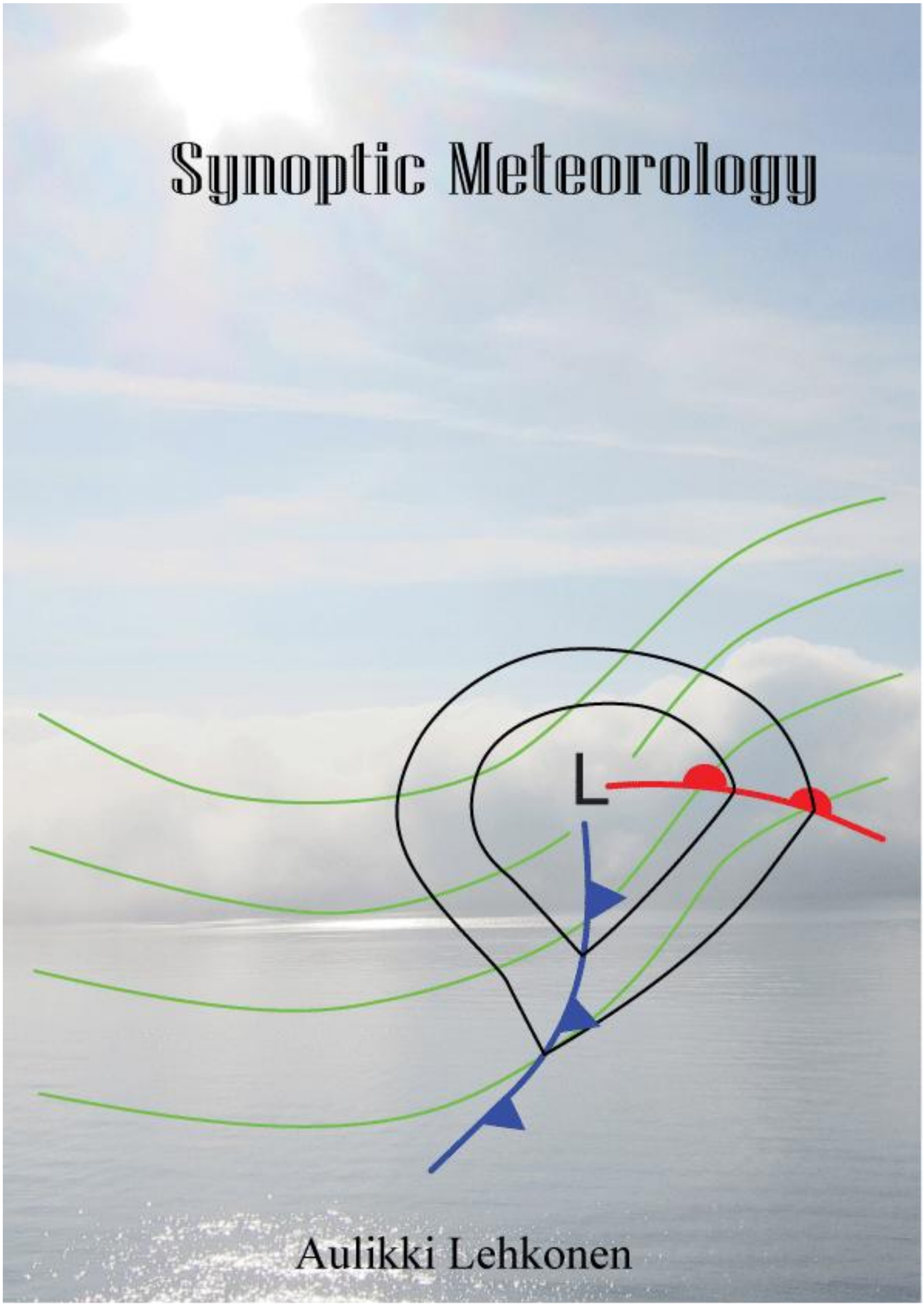


Eumetrain

Contents

1. Introduction

1.1. What is synoptics?

- ”synopsis”: forming a broad but robust picture of the general state of the atmosphere

- in synoptics, all information concerning the state of the troposphere is taken into account: observations and the parameters produced by numerical models.

- synoptics will help a meteorologist understand the state of the troposphere; what is happening and why, and what might be taking place in the near future

The theoretical basis of synoptics comprises atmospheric dynamics and mean flows.

Synoptic examination utilizes parameters calculated with the laws of thermodynamics and fluid dynamics.
—> One has to understand the equations derived from them and their practical significance!

Required tools and concepts of dynamics (Martin, chapters 1-6)

- vector calculus (vector components, nabla, cross product, partial derivatives)
- advection, divergence, vorticity, potential vorticity, deformation
- the basic forces: pressure gradient, gravitational, centrifugal and Coriolis forces
- the equations of motion, the conservation of momentum and mass
- hydrostatic equilibrium
- geostrophic and ageostrophic winds
- potential temperature
- thermal wind
- gradient flow
- barotropic and baroclinic atmospheres
- the omega equation and Sutcliffe Development Theory

1.2. The synoptic scale

- Horizontal dimensions of 500-5000 km
- Duration around 1-10 days
- Rossby number < 0,1

$R_o = U/L\ f_o$

Weather systems by size:

Global scale (5000 – 10 000 km)
- time scale 5 days – permanent
- e.g. polar vortices, subtropical highs
- satellite images, radar combinations

Synoptic scale (300 – 5000 km)
- duration 1-10 days
- e.g. mid-latitude cyclones and anticyclones
- soundings, radar images, satellite images

Mesoscale (3 – 300 km)
- duration 0,5 – 24 h
- e.g. sea breezes, thunderstorms
- synoptic observations, radar images, large resolution satellite images

Microscale (1 mm – 3 km)
- e.g. sea breeze, formation of hail
- radar images, measurements of individual parameters

However, it should be noted that:

- The classification is rough and not always clear; for example, a jet stream may be 5000 km long and 300 km wide

- Planetary waves also belong to the domain of synoptic observations, as they regulate the development and motion of mid-latitude pressure lows.

- Synoptics also deals with mesoscale phenomena like thunder. However, in such cases the focus is not on the structure of an individual cloud, but rather on the presence of thunder as part of a larger weather system or in a specific synoptic environment.

Large scale systems combine those components that cause mesoscale weather.

For example, the formation of thunder requires humidity, instability and buoyancy; if two of these factors are already present, adding the third will increase the chance of thunder.

Onward from the synoptic scale to the mesoscale!

1.3. Analysis and interpretation

The tasks of a synoptic meteorologist:

- Analysis of weather observations
- Interpretation of numerical fields
- Combining the analyzed observations and numerical parameters with conceptual models

Synoptic interpretation of the state of the troposphere, a four-dimensional concept of the state of the weather:

simultaneous consideration of observations and numerical prediction fields

analysis —> the current state of the troposphere

diagnostic laws of physics —> understanding the state of the troposphere

prognostic laws of physics —> the future state of the troposphere

Observations:

Synop observations

- other observations: automatic, flight and road weather observations
- soundings
- satellite images
- radar images
- observations from airplanes

A lot of observational data goes straight into the initial conditions of models. In particular, information from satellites is used to patch up the sparse network of oceanic observations.

Synop observations are analyzed.

Analysis means

- visualizing the isolines of parameters
- defining the areas where a given phenomenon may occur
- identifying observational errors
- achieving an unambiguous interpretation of the data

An analysis will also suggest which numerical parameters are to be paid attention to!

Example: analyzing surface SYNOP observations

- Isolines of isobars
- Surface pressure tendency isolines or isallobars
- Local weather (areas with precipitation, fog/mist, thunder, showers, etc.)
- Wind convergence zones
- Isotherms
- Other things (as required); wildfires, volcanic ash clouds, sizable dust clouds, etc.

1.4. Conceptual Models

Conceptual models:

- simplified representations of the properties of weather systems
- are not necessarily unequivocal; weather systems are formed, die out, undergo change and are not always clear with regard to every parameter
- vary according to surface, season and time of day
- contain information about a system's development over time

There are manuals of conceptual models (SatManu, Lehkonen: käsitemallit), but interpreting the weather is subjective.

A sample of conceptual models:

Synoptic scale	Mesoscale
Tropopause jet streams, isotachs and altitude	
Pressure lows and highs	Weather phenomena
Temperature anomalies	Fog and Cb clouds
Fronts and troughs	Winds at boundary layer
Conveyor belts	

Why do we need manual analysis and interpretation?

- The identification of conceptual models cannot be automated with the existing equipment.
- Errors in the model may be identified, along with developments deviating from the model's predictions.
- Local conditions and mesoscale phenomena (e.g. surface, season, history, etc.) can be taken into account.
- People are better than computer programs at inferring observational errors and the effects of sub-synoptic scale weather and environmental phenomena.
- It has been observed that the drawing of lines and conceptual models on a map improves a person's understanding of the subject more than simply looking at finished products. On the other hand, synoptic charts contain a lot of information in condensed form, and when people conduct analysis on their own, they tend to examine things more closely.

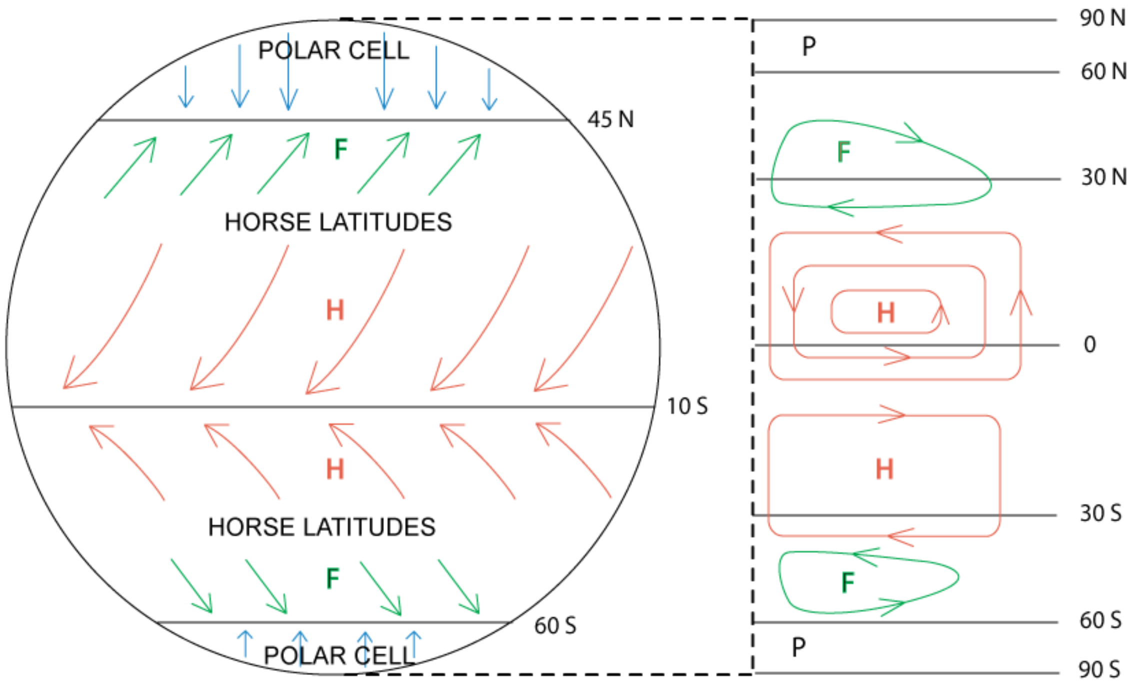
An analysis chart may be

- A practical tool for oneself and one's colleagues, like surface maps
- A customer product
  - Newspaper weather maps
  - Aviation briefing maps

2. Tropospheric circulation

2.1. Mean wind conditions globally

A schematic of Earth's mean winds:



The polar cell is very weak, so it does not show in the vertical cross section. In summer the circulation moves northward; for example, because of the warming of the large continents in the northern hemisphere, the intertropical convergence zone (ITCZ) lies approximately at the 15^{th} latitude north, generating the Indian monsoons as well as causing the Hadley cell of the southern hemisphere to be much stronger than that of the northern.

Tropical and extra-tropical: general circulation

The temperature difference between the tropics and the poles is the most important factor behind general circulation. The temperature difference is largest in winter, when the circulation is also at its strongest.

Atmospheric circulation in the tropics (0 – 30)º is dominated by a Hadley cell, which

- transfers dry air (energy) toward the poles
- transfers latent heat toward the equator
- transfers angular momentum upward
- produces kinetic energy (→ subtropical jet stream)

Outside the tropics (30 – 90)º changes in the upper waves

- transfer dry energy and latent heat toward the poles
- transfer impulse momentum from the equator toward the poles
- produce the majority of kinetic energy of the circulation
- determine the weather in the area in question

“Disturbance” or “weather disturbance”: the birth and strengthening of an upper wave (a wave in the mid- and upper troposphere pressure fields), which results in the formation of a mid-latitude cyclone. They are accompanied by wind, rains and temperature changes. There are 4-5 disturbances present at all times to maintain the thermal equilibrium between the mid- and high latitudes. Instead of mean flows, synoptics concentrates on transient weather, which arises mostly from the disturbances.

Note: in general a “cyclone” is a counterclockwise vortex; however, in synoptics it refers to a mid-latitude moving low, and not for example a small polar low or a tropical hurricane.

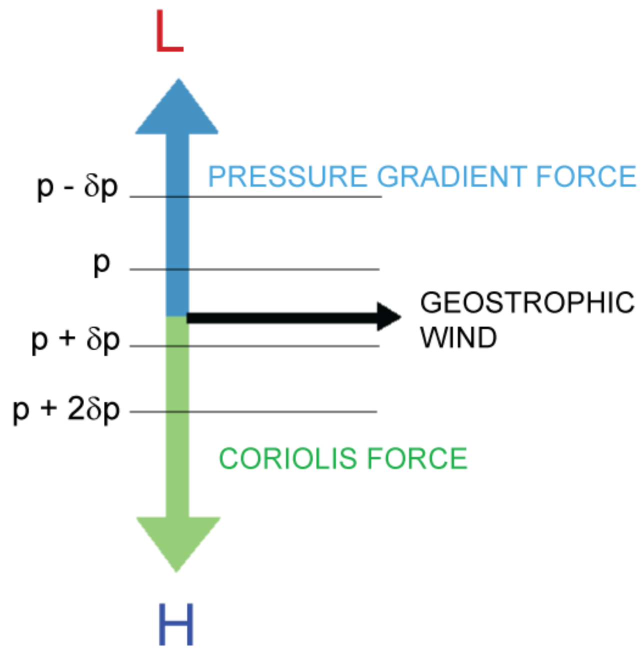
Note that the often used 'westerlies zone' is a rather vague concept; in meteorology the preferred terms are air masses and mid-latitudes.

2.2. Geostrophic and thermal winds

Geostrophic wind (Martin p. 61- 62), MetED 'Topics in Dynamic Meteorology: Thermal Wind'

The equations describing the synoptic scale currents of the mid-latitudes have two terms that are at least an order of magnitude larger than the others: the pressure gradient force and the Coriolis force. This being so, it can be assumed that in a state of equilibrium the two forces counter each other; the pressure gradient force is equal and opposite to the Coriolis force.

In the northern hemisphere:



$$\overline{V}_g = \frac{1}{\rho f} \hat{k} \times \nabla p$$

Geostrophic winds blow parallel to the isobars. The tighter the pressure gradient, the stronger the geostrophic wind!

Geostrophic wind represents the real wind with an error margin of 10-15 % everywhere else except in the boundary layer, the jet streams and wave disturbances associated with them, and the tropics and polar regions.

Geostrophic equilibrium only applies to large-scale flows; for phenomena on a small scale (sea breeze, tornados, Cb clusters, etc.) the pressure gradient force is much stronger than the Coriolis force.

According to the geostrophic wind law, where westerlies prevail there must be high pressure near the equator and low pressure near the poles. This is seen in mean pressure distributions: the Horse latitude highs and polar vortices.

Thermal wind (Martin: p. 89-93)

A warm column of air extends higher than a cold one. That is to say, in cold air each pressure level lies lower down than in warm air. If the columns are brought side by side, the pressure levels slope downwards toward the cold air. A horizontal geopotential gradient forms, and it is proportional to the pressure gradient. The magnitude of the geopotential gradient increases with height.

Thus, in geostrophic conditions geostrophic winds also increase with height.

—>

Horizontal temperature gradient and the vertical shear of geostrophic wind are mutually dependent.

vertical shear of geostrophic wind = thermal wind

On average, hydrostatic equilibrium and geostrophy provide a good representation of the troposphere, and these assumptions apply to most of it. From them we can derive the thermal wind law:

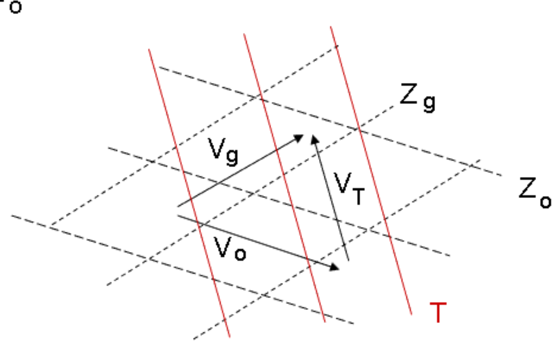
$$\frac{\partial \overline{V}_g}{\partial p} = -\frac{R}{f\,p}\hat{k} \times \nabla T$$

The thermal wind law explains why westerlies prevail in the mid-latitudes.

The vertical shear of geostrophic wind (i.e. thermal wind) is parallel to the isotherms.

A vector representation of a thermal wind:

$$\overline{V}_T = \overline{V}_g - \overline{V}_o$$



2.3. Upper waves

Upper waves determine the weather outside the tropics.

The basic principle is that orography and the uneven distribution of thermal radiation between the equator and the poles generate an upper flow, in which waves (weather disturbances) are constantly forming.

In accordance with the geostrophic wind law, where westerlies prevail there are, in the mid- and upper troposphere, areas of high pressure near the equator and areas of low pressure near the poles (polar vortices).

Zonal index

The zonal index is a measure for the strength of mid-latitude westerlies. It is expressed either as a horizontal pressure difference between latitudes 35 and 55 or as the corresponding geostrophic wind.

High zonal index (over 8 hPa): strong westerly component, zonal flow. There are short, fast and short-lived waves in the flow. Isotherms and isohypses have differences in phase and amplitude. Weather changes rapidly.

Low zonal index: weak westerly component, meridional flow. There are long, slow and persistent waves. Isotherms are almost parallel to isohypses. Weather changes slowly.

Index circulation: when the index is low, a westerly flow will turn more and more meridional as the troughs and ridges strengthen and finally tear off from the main flow into separate vortices (cyclones and anticyclones). After this the current will shift rapidly into a high-index state. The length of a cycle is 3-8 weeks, although with especially high and low indices the circulation is faster.

Wavelength and movements of the upper waves affects the type of weather: will a given air mass stay where it is, or are the conditions undergoing rapid changes (high or low zonal index).

In a westerly flow, a wave may travel east (progression) or west (retrogression) or remain still. This is determined by the wave's velocity, wavelength and latitude, and it is represented by the so-called β-parameter, or Rossby parameter.

β-parameter (Rossby 1939) :

$$\beta = \frac{2\Omega\cos\varphi}{a}$$

Ω = Earth's angular velocity
a = Earth's radius
φ = latitude

U = The velocity of the westerly flow (basic flow), or Rossby velocity (assumed constant)
c = phase velocity of a wave moving in the basic flow
L = wavelength of a wave moving in the basic flow
k = wave number, $2\pi/L$

$$c = U - \frac{\beta}{k^2}$$

progression (wave travels along the basic flow): $c > 0$
retrogression (wave travels against the basic flow): $c < 0$

3. Air masses

3.1. Defining an air mass

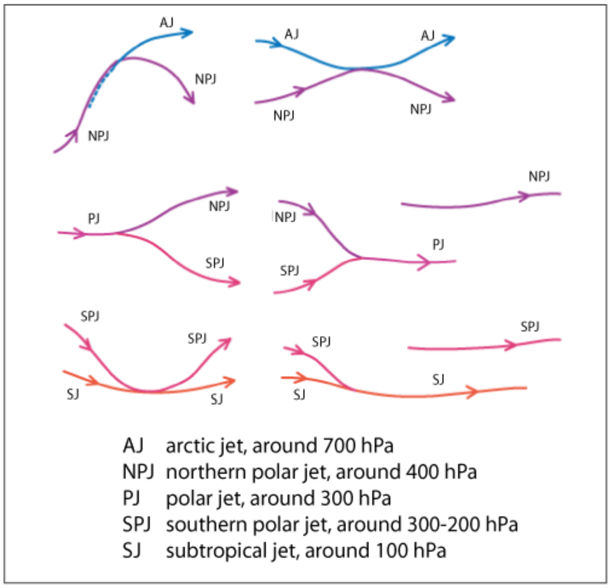
Air mass:

- A large-scale volume of air
- Fairly homogenous in horizontal temperature and humidity
- Vertical variations in temperature and humidity almost identical throughout the volume
- Remains or travels above a certain region of homogenous conditions long enough to reach a relative equilibrium with the surface
- The properties of the source region follow from the radiation balance of the surface and air.
- Undergoes certain changes as it drifts away from the source region

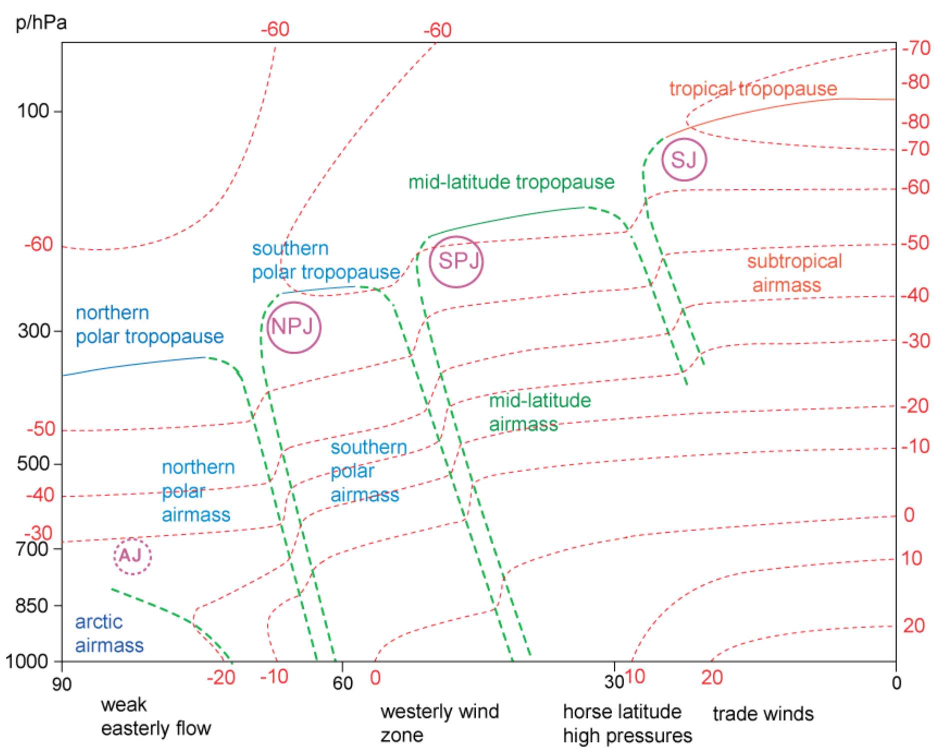
Between air massses there are baroclinic zones where temperature and humidity change rapidly as one crosses the zone.There is always a jet stream above the baroclinic zones, near the tropopause. Baroclinic zones are where fronts are formed.

Air masses and the jet streams associated with them are generally divided into tropical, subtropical, mid-latitude, polar and arctic air masses. Polar air masses can be further divided into southern and northern masses where the polar jet splits into northern and southern branches.

Jet streams and their branches occurring in Finland:



The air masses and associated jet streams in the northern hemisphere are seen in this cross section of the troposphere:



3.2. Classification of air masses

Under the most commonly used system of classifying air masses, their abbreviations take the form xY, where

x = c (continental) or m (maritime)
Y = T (tropical), P (polar) or A (arctic)

Another class, ML (mid-latitudes), is used in Finland, but it is rather vague, as it is constantly transforming and it does not have a homogenous source region. In Finland the polar air mass is divided in two parts wheneve such a distinction can be made in the jets: NP (northern polar) and SP(southern polar). The air masses used in Finland are therefore A, P, NP, SP and ML. The tropical air masses never reach Finland.

The original Bergeron classification system from 1928 also contained a term which indicated whether the air is warmer or colder than the surface.
There is also an E (equatorial) class, which is occasionally used in the United States.

Source regions of air masses

"an area of homogenous conditions, where the air mass lingers" e.g. sea, ice, tundra, steppe, desert, plateau

Source regions:

- Arctic air mass: surface covered with snow and ice, with a strongly negative radiation balance
- Polar air mass: polar and other high-latitude regions
- Mid-latitude air mass: the poleward borders of horse latitude highs
- Tropical air mass: trade wind zone

3.3. Properties of air masses

When defining an air mass, the following factors are taken into account:

- the altitude, temperature and shape of the troposphere
- location of jet streams relative to each other; which jets are present?
- location of jet streams relative to fronts (more on this in chapter 6)
- season
- T850 hPa and T500 hPa relative to statistical mean for that time
- mean temperature (thickness) between two pressure surfaces

- humidity of the surface layer

There have been no precise definitions for the treshold values of the temperature in different air masses. There is an exception in Finland, where an air mass is classified as arctic when T(850 hPa) < -18 °C.

Note that temperature at 2 meters says very little, as the effect of the surface is considerable.

Temperature and thickness of an air mass

The temperature of an air mass is best represented by the temperature at 850 hPa, or the dry or moist potential temperature at the same altitude. (In mountainous regions 700 hPa is used instead of 850 hPa)

The mean temperature or thickness between the pressure surfaces is also a good representation of an air mass:

$$dz = -\frac{dp}{\rho g} = -\frac{dpRT}{pg}$$

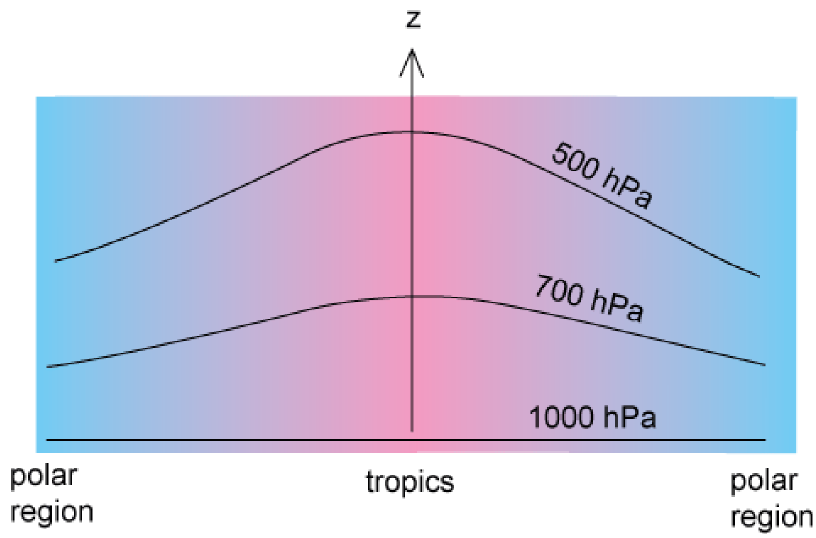
$$\Rightarrow$$

$$h = \frac{R\overline{T}}{g}\ln\frac{p_o}{p}$$ where

ρ = thickness
R = gas constant
p and p_o = the pressure surfaces whose intervening thickness is being calculated
$\overline{T}$ = mean temperature of the layer
h = height of the layer.

Thickness is calculated from layers (850-500) hPa, (1000-850) hPa or (1000-500) hPa.

The warmer the layer, the higher it reaches:



Aspects to note about air masses:

- Cold air usually arrives in Finland from the north and warm air from the south, but an air mass is not generally determined by where it comes from; in Finland, for example, a northeasterly flow may carry with it either cold or warm air in summer.

- Vertical motions are also a crucial part of air mass movements. For example, on the east side of a high pressure ridge cool air flows to the south descending, while on the west side warm air flows to the north and ascends. As the ridge travels east, the lower and upper parts of the troposphere may hold different air masses in the same place for some time.

- The weather within a homogenous air mass is determined by the properties of the surface and the air mass itself. In the zones between them, weather is determined by synoptic scale weather systems - often fronts.

Identifying air masses from soundings

Height and temperature of the troposphere (approximate):

- In polar air: altitude 350-250 hPa, temperature -50…-60 °C. Shape fairly undefined or sharp, in which case the lower stratosphere is almost isothermic.

- In mid-latitude air: altitude 200-100 hPa, temperature around -60 °C. Shape often sharp, in which case temperature increases with altitude in the lower stratosphere.

- In tropical air twofold: the lower one is around 200 hPa and -60…-70 °C, while the upper is < 100 hPa, -70…-80 °C.

- The arctic air mass is limited to the lowest part of the troposphere, in which case there is a polar air mass above it.

However, when it comes to identifying air masses, it is the jet streams that define them; soundings are usually more vague.

Finnish weather in different air masses

Cold air masses, A and PP

- The tendency to labilize quickly initiates convection
- Visibility is good, except when reduced by hydrometeors or lithometeors
- At sea the temperature at 2 m is lower than the sea surface temperature
- Gusty wind

Summer:

- On land there are convective clouds in during the day and clear skies at night
- At sea there are convective clouds, particularly at night in late summer
- In deep convection there are local showers and thunder in the afternoon and evening
- Close to ground the diurnal variation in temperature and humidity is large (in spring as much as 20 °C)

Winter:

- Large raindrops, large dendrites (snowflakes) or graupel below -5 °C

- Flowing snow in gusty wind
- At sea there is a strong upward flux of water vapor, as well as sea smoke when air temperature is below -18 °C
- Strong surface inversions in weak wind conditions

Note that when inversion is strong, in Finland usually in winter, the strengthening or subsiding of the wind often causes a greater (even opposite) change in the temperature than the horisontal temperature advection.

Rule of thumb:

> An air mass is "cold" when T850 < -10 °C in winter and < 0 °C in summer. Correspondingly, T500 < -30 °C and < -20 °C.

Warm air masses SP and ML

- Wind is steady
- Visibility often poor (haze or mist)
- There can be radiation fog on the continent especially in spring and autumn, and advection fog in spring either at sea or over melting snow.

Summer:

- Weak surface inversion at night
- When the air is dry in the summer, there are shallow Cu clouds or clear skies during the day
- In potentially unstable situatons there can be showers and thunderstorms in the late afternoon and evening, provided that the air is humid enough

Winter:

- Often there are St/Sc clouds, drizzle, snow grains or weak snowfall.
- When the skies are clear, the lowest layer of the troposphere cools gradually, and the inversion remains throughout the day.
- Inversion situations are accompanied by fog or mist when the temperature of a wet surface is around zero.

Phase of precipitation

Thickness can be used to predict the phase of precipitation. In Finland, when the 1000-850 hPa thickness is below 1275 m, precipitation occurs as snow, and when it is above 1310, it occurs as water. The probability of sleet is greatest when the thickness is 1295 m. (Hankimo 1969)

The phase product of Finnish Meteorological Institute radar images is derived from a statistical equation based on 150,000 SYNOP observations:

$$P = \frac{1}{1 + e^{(22-2,7T-0,2RH)}}$$

Here P = probability of liquid precipitation (P > 0,7 means water and P < 0,3 means snow), T = temperature and RH = relative humidity.

Effect of relative humidity: in dry air snow showers may occur even at +8°C, but in humid air, for example below a Nimbostratus, precipitation becomes water at +2 °C at latest.

3.4. Transformation of air masses

An air mass changes as it moves away from its source region.

Processes that can transform an air mass:

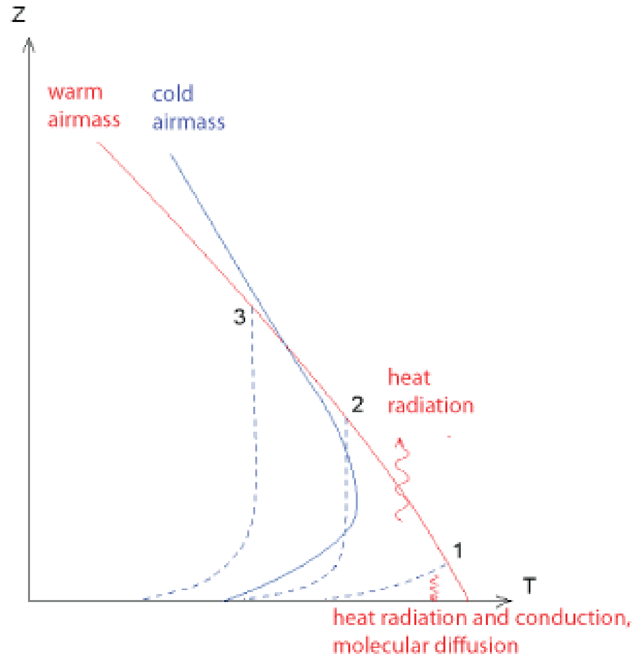
- The cooling effect of the surface
- The warming effect of the surface
- Synotic scale ascent + release of latent heat
- Synoptic scale subsidence
- Turbulent mixing

1. The cooling effect of the surface

A relatively warm air mass arrives over a cold surface while cloudiness and windiness are low or decreasing.

Heat is transferred from the air into the surface through radiation, molecular diffusion and, in the boundary layer (< 1 mm from the surface) conduction.

Initially this causes a very low but strong surface inversion. This stabilization of the air dampens turbulent mixing particularly if the wind is calm. The layer above the surface undergoes radiative cooling while the surface inversion becomes stronger and stronger. In Finland the radiative cooling effect of the surface is particularly strong over snowy ground in the spring and at sea in early summer.

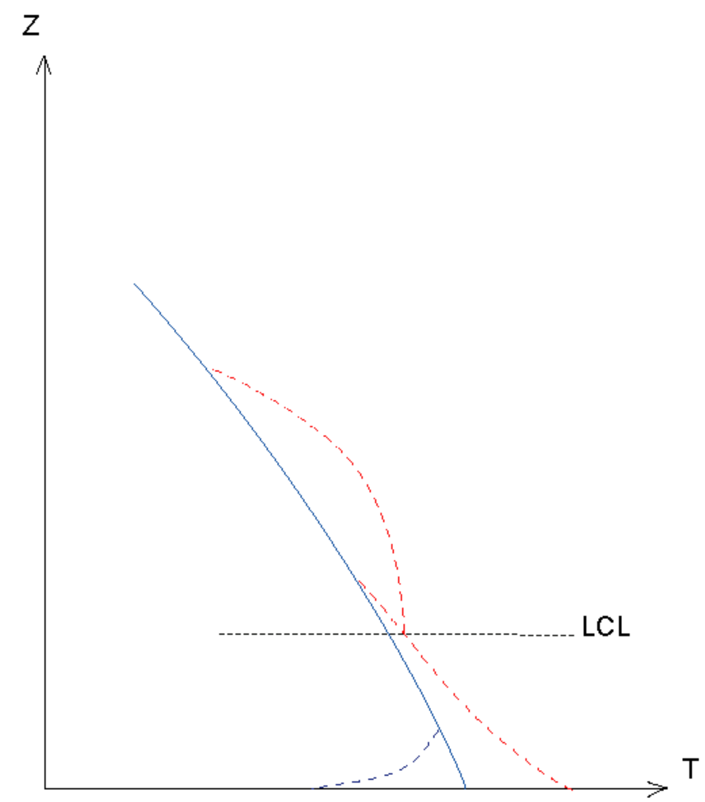


2. The warming effect of the surface

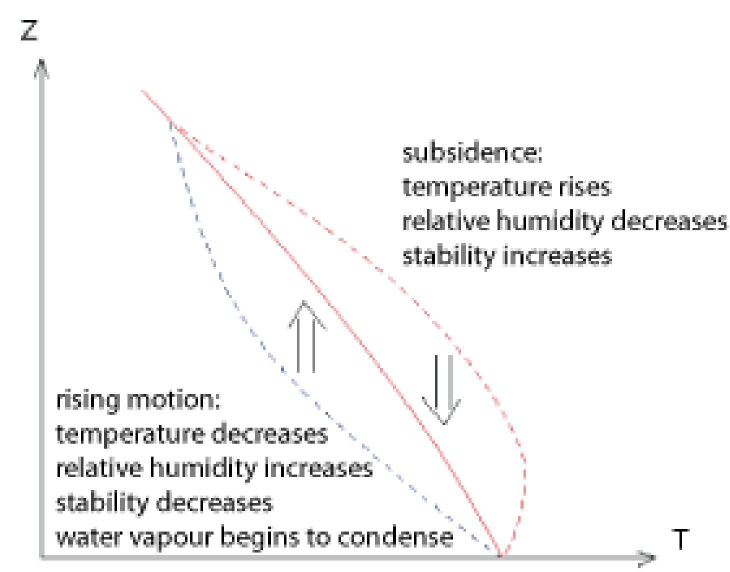
Conduction, radiation and molecular diffusion cause the air layer to labilize over a warm surface, which initiates convection. The heating caused by the convection often reaches far higher upwards than the cooling caused by the surface.

If Cb clouds are formed in the convection, the release of latent heat warms the mid- and upper parts of the troposphere above the lifted condensation level (LCL).

Oceans are very effective at warming the lower troposphere because seas have a near constant surface temperature on the synoptic time scale.
In Finland, the warming of cold air masses is strongest in June-July over land and in August-January at sea.



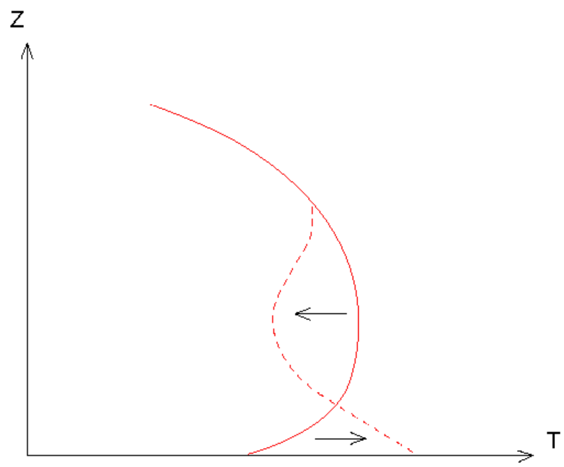
3. Vertical motions on the synoptic scale



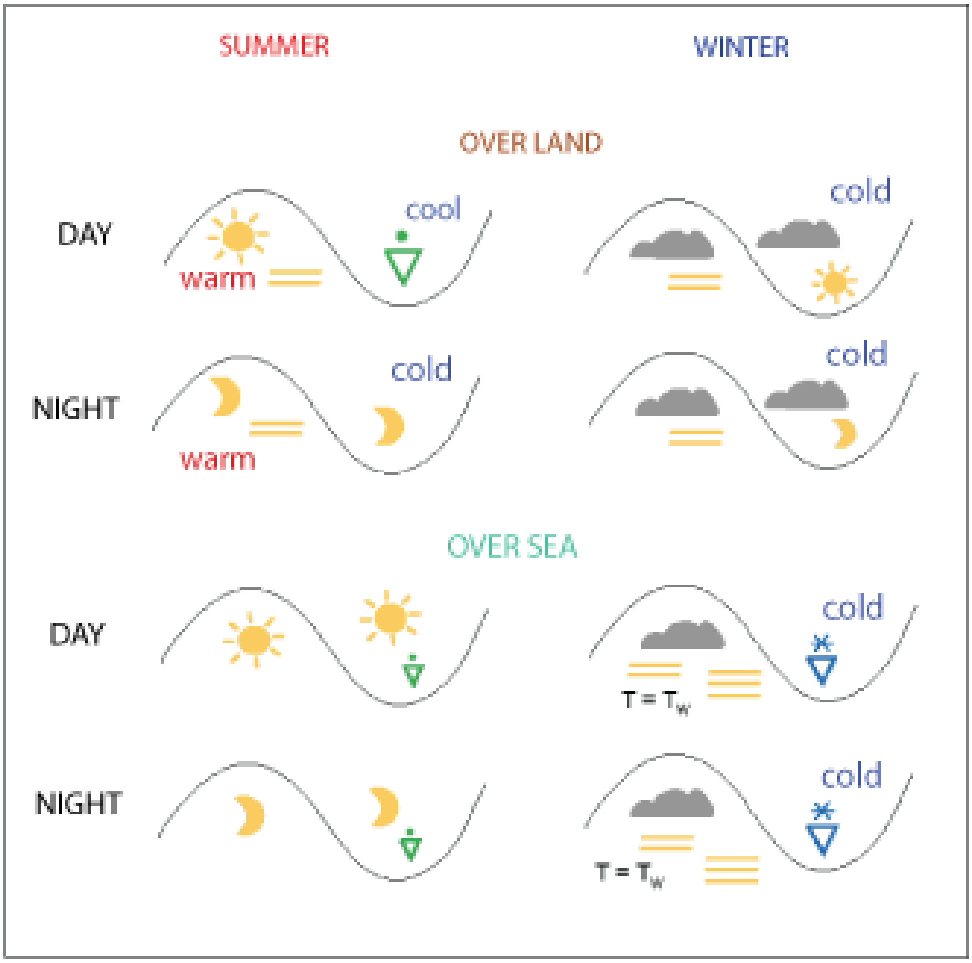
Note that convection can occur over a warm surface even if there is weak descending motion on a larger scale. Typically this occurs in weak high pressures in summer. If the air is humid enough, showers and even thunder can occur in the afternoon and evening.

4. The effects of turbulence and mixing

Turbulence is caused by convection and friction
In the boundary layer the air is well-mixed and the potential temperature is almost constant.



The weather in Finland predominantly adheres to the following pattern in 500 hPa highs and lows, when there are no high pressure gradients or weather disturbances nearby:



This is a rough estimate; there are many other factors affecting the weather. For example, a high can be accompanied by afternoon showers in the summer, a high can initially have clear skies in winter, an inversion can remain in a cold air mass throughout the day in winter, etc.

3.5. Air mass movements

Martin p. 123-130, esp. 129-130

Air mass movements can be examined with the help of potential vorticity (PV): Potential vorticity is a concept corresponding to potential temperature; a measure of potential for vorticity in an air parcel.

The usability of this parameter with large air volume movements is based on its conservation in adiabatic flows:

$$PV = \frac{\zeta + f}{\delta p} = \text{constant}$$

, where

where ζ = relative vorticity
f = the Coriolis parameter
δp = the thickness of the column of air.

Meridional movements

Let us consider the movement of a parcel of air on a north-south axis, assuming the flow is adiabatic and that vorticity is low. This is true on a large scale.

In a situation where $\zeta = 0$, δp only depends on f. In such a case the column of air is compressed adiabatically as it moves toward the equator. As the air column compresses, its temperature rises.

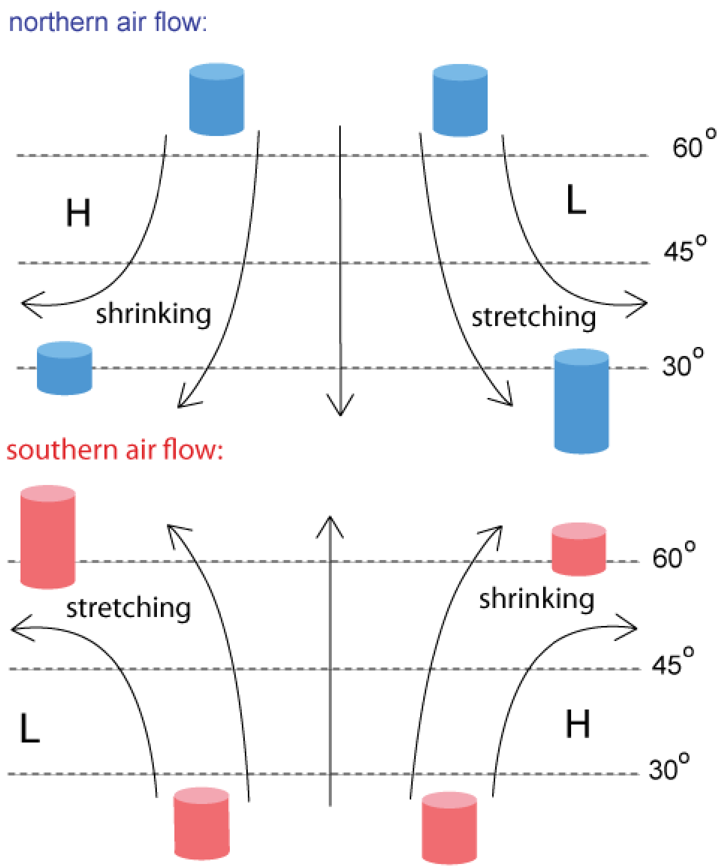
Consequently, the air mass changes as it drifts sufficiently far south. The colder the air was originally, the farther it can travel without losing its temperature difference relative to the surroundings. According to Palmen & Newton (1969), observations suggest that polar air cannot travel beyond 25° latitude.

In addition to adiabatic heating, the air is warmed by the surface. Heat transfer is greatest when cold air is moving toward the equator over oceans, which have large heat capacities, and smallest over a snow-covered surface. Cold air can thus travel farther south over a snow-covered continent than over an ice-free sea.

As an air parcel moves toward the equator, cyclonic vorticity increases and the parcel curves east.

Accordingly, if a polar air mass has been advected to the mid-latitudes, it has to contain cyclonic vorticity.

As an air parcel moves polewards, it stretches vertically and cools unless its relative vorticity decreases considerably. If the vorticity is anticyclonic, the parcel's depth has to be small.



Note:

- The difference between cyclonic and anticyclonic air masses is greater than that between warm and cold ones!
- Deep convection does not occur in anticyclonic flows.

When an air mass curves anticyclonically in a strong cold air flow, it contains subsidence, the upper troposphere warms and the air stabilizes. In such a case there will be no deep convection. If the surface is very warm, an anticyclonic flow can also generate low convection (Cumulus Humilis), but not thunderclouds. In a cyclonic flow the air mass becomes more labile and deep convection does occur.

When air moves toward the equator, the ground heats its lower parts concurrently with the adiabatic heating taking place above. The heat transfer is greatest when the air travels over an ice-free ocean, and weakest over a snow-covered surface.

Accordingly, warm air moving polewards is slowest to transform when it travels over a continent in summer or over a sea in winter.

In summer, the temperature in Finland is highest when there is a flow from the southeast. In winter, the mildest weather follows winds from the west-southwest.

The coldest episodes in Finland are related to northerly flows in summer and east-northeasterly flows in winter.

Zonal movements

Chapter 2.2. dealt with upper waves. Weather in Finland is determined by the movements of these waves.

When the zonal index is high and the basic flow fast, it takes 0,5 – 2 days for a wave to pass. A large slow wave, on the other hand, can persist for up to several weeks.

The large scale weather:

- There is changeable weather near the flow maximum (near the high pressure gradient zone), where weather disturbances move and air masses change rapidly.

- A large trough is accompanied by a polar air mass, and the weather is cool.
- A large ridge is accompanied by warm air, dry in summer and humid in winter.

Discontinuous retrogression of the upper waves

Sometimes a trough or ridge moves west, but these cases are rare. Rather, it is more common for a large trough or ridge to appear to stop and move westwards, which is a case of discontinuous retrogression. This phenomenon is related to an upper flow changing from zonal to meridional.

In an adiabatic, homogenous flow on an isentropic surface, the potential vorticity is preserved:

$$\frac{\zeta + f}{\delta z} = \text{constant}$$

Unless the height z changes, absolute vorticity $\zeta + f$ is constant, so when f decreases as one approaches the equator, relative vorticity ζ increases, meaning the flow becomes more cyclonic.

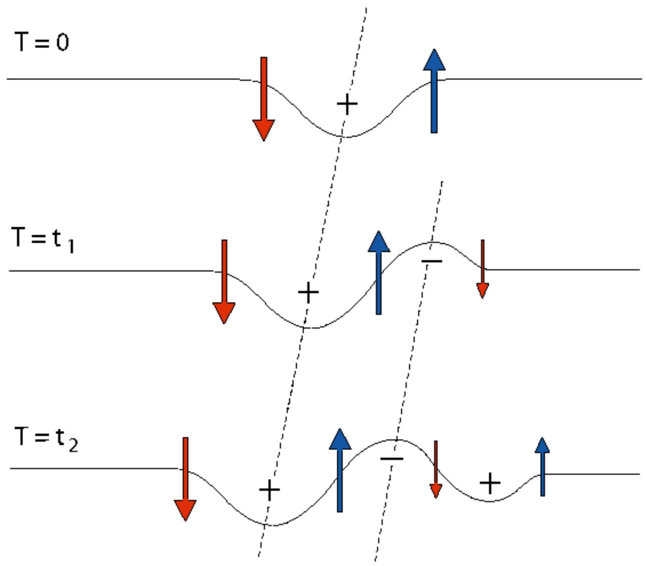
Vorticity is given a closer look in chapter 5. Here it is only used to describe the movement of a trough's bottom/ridge's top.

The flow gathers cyclonic vorticity. This is also known as a positive vorticity anomaly. In a westerly basic flow this anomaly is accompanied by counter-clockwise circulation, which advects positive vorticity to the west of the anomaly, and negative vorticity to the east of it.

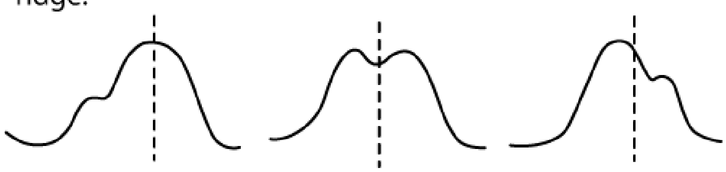
Retrogression is related to weather that prevails for a long time.
For example, in Finland in summer the weather stays cool and rainy so long as there is a large trough of low pressure in western Europe. Small troughs travel along the trough's easterly borders from the south to the northeast. The cold, rainy weather persists as long as the axis of the larger trough stays still or returns west instead of continuing to the east.

Retrogression of the upper waves:

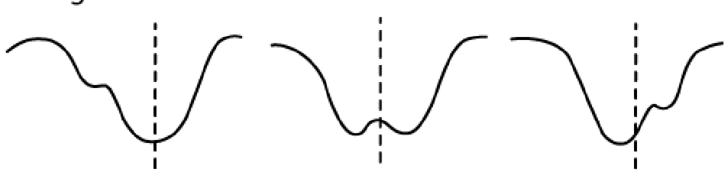
Red arrows: positive vorticity. Blue arrow: negative vorticity.



ridge:



trough:



4. Boundary layer and weather

4.1. Physical properties of the boundary layer

The boundary layer is the lowest layer of the troposphere. Above it lies the so-called free atmosphere. Flows in the boundary layer are affected by the surface. They are also turbulent, and the geostrophic approximation does not apply in the boundary layer.

The thickness of the boundary layer is influenced by:

- roughness of the terrain
- time of day (amount of incoming radiation)
- season (surface temperature, amount of incoming radiation)
- stability of the air mass

Weather phenomena in the boundary layer:

- fog and mist
- stratus and stratocumulus clouds
- strong temperature inversions
- gusts of wind
- anabatic winds and sea breezes

Properties of a turbulent flow:

- rapid fluctuation
- describing random changes requires statistical methods
- continuous flow: even the smallest vortices are much larger than the free distance between molecules
- the formation of vortices requires energy, which comes from the basic flow
- it is impossible to predict the details of vortices, but their collective influence can be described by parameterization

The flow mixes

- momentum
- temperature
- humidity
- trace gases and particles

2-m winds are turbulent, meaning changes take place rapidly there. This is why SYNOP observations give the 10-minute mean wind (both velocity and direction). The parts of the boundary layer:

1. Surface layer

    - the lowest 10 % of the boundary layer
    - friction turns the wind towards lower pressure around 20-30°on land and 12-17°at sea, on average.
    - the mean wind is roughly 50 % as strong as the upper wind on land and 70 % as strong at sea

2. The Ekman (or spiral) layer

    - Wind strengthens and veers with altitude. On the top of the boundary layer the wind becomes almost geostrophic.

    - Sc clouds and fog often form in the upper and lower parts of the boundary layer, respectively.

Ekman spiral

The so-called Ekman spiral is a simplified mathematical description of the winds in the Ekman layer. Assumptions are that the density and viscosity in the layer are constant, motion is horizontal and steady, isobars are straight and parallel, and the basic flow is constant throughout the layer.

The concept of the Ekman spiral was developed by the Swedish oceanographer Ekman in 1902 to describe ocean currents. The concept was applied to the troposphere six years later.

The components of the Ekman wind are:

U parallel to the pressure gradient
V perpendicular to the pressure gradient

$$U = - G\, e^{-\beta} \sin\beta$$

$$V = G(1 - e^{-\beta} \cos\beta)$$, where

G = geostrophic wind velocity (basic flow)
β = z(f/2K)(exp)1/2
f = Coriolis parameter
z = height
K = eddy viscosity, which increases the internal friction of the flow by transferring the momentum of the turbulent vortices to it

The level H where U=0, (the real wind is parallel to the geostrophic wind) is called the geostrophic wind level or the gardient wind level. Its height H is:

$$H = (2K_M / f)^{1/2}(3/4\pi + \alpha_o)$$

where $\alpha =$ the angle between the surface wind and surface isobars

At this altitude the real wind speed is a little greater than that of the geostrophic wind, depending on the value of β .

In other words, wind in the Ekman layer depends on the Coriolis force. It grows exponentially with height.

An interactive module for exploring winds in the boundary layer according to elevation and roughness of the terrain: http://ww2010.atmos.uiuc.edu/(Gh)/guides/mtr/fw/bndy.rxm

Inversion

When the radiation balance is negative (at night, in winter), the layer near the surface cools. This causes formation of an inversion where temperature increases with height. In Finland, surface inversions of up to 30 ºC may occur in winter if the skies are clear and the air cold enough.

Under inversion conditions, the boundary layer is also called the inversion layer.

In strong surface inversions the wind is calm. Turbulence and mixing cease. This leads to the accumulation of air pollutants in cities and industrial areas, at worst to smog.

4.2. Diurnal variation of the boundary layer

The following factors increase turbulence in the boundary layer:

- rough terrain
- strong upper wind

$\longrightarrow$ mechanical turbulence

thermal radiation from the surface in summertime (unfrozen ground)

$\longrightarrow$ thermal turbulence

Thermal turbulence subsides at night, and hydrostatic stability weakens the mechanical turbulence. Winds become weaker and more steady, and the boundary layer grows thinner.

In Finland the height of the boundary layer in summer is between 1,5-2,5 km in daytime and some hundreds of meters at night. The diurnal variation is smaller in winter because there is no thermal turbulence. The height of the boundary layer in winter is 0-1,5 km.

Diurnal variation in a convective boundary layer in summer:

If the wind is weak and the surface warm during the day, the layer just above the ground is statically unstabile. Turbulent heat flux is strong close to the surface and diminishes linearly with height. There are a lot of vortices (including large ones) in the Ekman layer, and it is considered "well-mixed": the mean parameters are near constant throughout the layer.

The height of the boundary layer is clearly defined and grows over the course of the day as turbulent vortices and thermals bring warm air upwards from the surface.

When the sun sets, thermal mixing and thermals cease. The surface cools, and first the surface layer and then the lowest part of the Ekman layer stabilize. The gustiness and mean velocity of winds drop rapidly. A so-called residual layer remains in the upper parts of the boundary layer for a time, sometimes even overnight, but turbulence is weak also there during the night.

In the following schematics there are examples of typical vertical profiles for potential temperature, winds, mixing ratio of water vapor, and pollutant concentration (Stull, 1988) during the day:

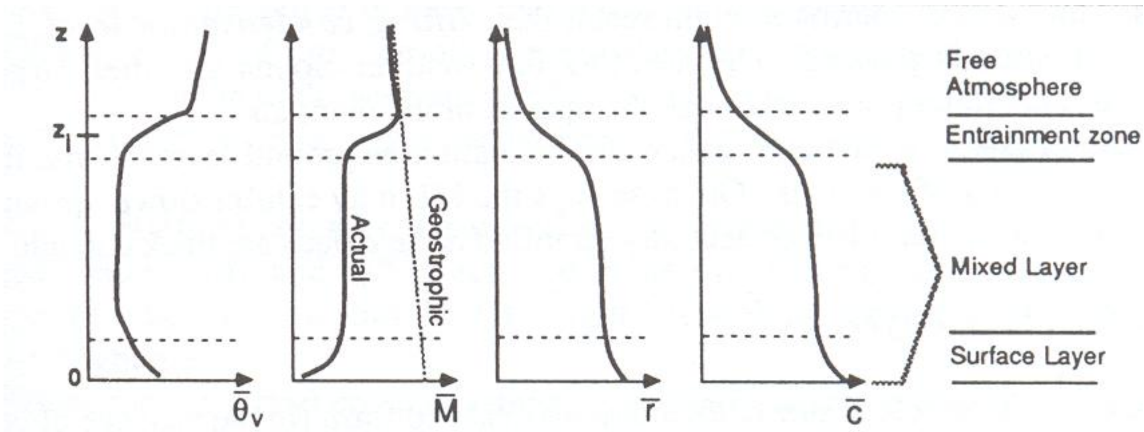


Fig. 1.9 Typical daytime profiles of mean virtual potential temperature $\overline{\theta}_v$, wind speed $\overline{M}$ (where $\overline{M}^2 = \overline{u}^2 + \overline{v}^2$), water vapor mixing ratio $\overline{r}$, and pollutant concentration $\overline{c}$.

Diurnal variation of a convective boundary layer (Stull, 1988):

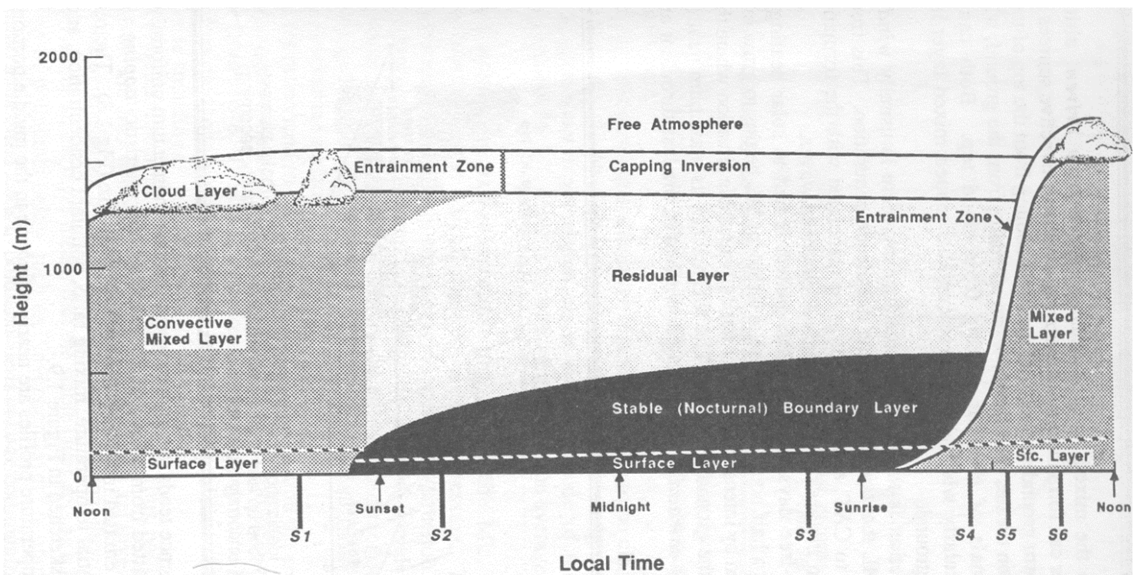


Fig. 1.7 The boundary layer in high pressure regions over land consists of three major parts: a very turbulent mixed layer; a less-turbulent residual layer containing former mixed-layer air; and a nocturnal stable boundary layer of sporadic turbulence. The mixed layer can be subdivided into a cloud layer and a subcloud layer. Time markers indicated by S1-S6 will be used in Fig. 1.12.

Entrainment

Consecutive rising bubbles of air intrude into the upper inversion, resulting in the mixing of air in the inversion layer and in the boundary layer. This is known as entrainment, and it causes a negative turbulent heat flux into the upper regions of the well-mixed layer.

Stabile boundary layer

A slightly stabile boundary layer is the most common kind in Finland. The turbulence in such conditions is (depending on the wind and stability) moderate or weak, as is mixing, and the wind keeps to a spiral structure in a shallow boundary layer.

Mean winds often have a slight maximum in the upper parts of the boundary layer. If the wind's speed exceeds that of the geostrophic wind by at least 2 m/s, the term used for it is low-level jet.

A stabile boundary layer may consist of several layers with minimal interaction.

Cold air pools

In mountainous terrain, heavy, stabile surface air that has cooled through radiation on a relatively calm evening, will flow down the slopes as a katabatic wind. Weak katabatic winds can travel far even over flat terrain. They fill valleys and canyons with pockets of cold, still air known as cold air pools. The phenomenon is found especially in central Europe in the Alps, and it is known as valley fogs.

Maritime boundary layer compared to a continental one:

- Air is humid and there are more clouds
- The boundary layer is usually almost neutrally stratified: there is less of turbulent heat transfer and convection is less common
- The annual and diurnal cycles of surface temperature are smaller

Stability of the boundary layer over the Baltic Sea:

- In spring and early summer the sea is colder than the air, and the boundary layer is stabile
- In autumn and winter the sea is warmer than the air, and the boundary layer is very lightly stabile
- Over sea ice the boundary layer is usually stabile

4.3. The effects of the surface

1. Humidity (in summer)

   A humid surface stores and conducts heat well, and a part of the incoming heat is lost through evaporation.

   Diurnal variation in temperature is smaller over a humid surface than a dry one; nights are warmer and days cooler.

   extremes: sea / dry peat

   In summer, the evaporation of nocturnal fog releases moisture into the air over the course of the morning, and convective clouds form more easily over these areas than over a drier surface.

2. Vegetation (in summer)

   Evaporation is most effective in:

   - low vegetation during sunny weather (risk of frost in drained open marshland!)
   - forests in overcast weather (light comes in from multiple directions, which results in less shade)

3. Snow

   - snow cover reflects 40-95 % of the radiation
   - the thawing of snow and frozen ground takes up a lot of energy
   - snow is a sink for humidity when below 0 °C and a source of it when above 0 °C

4.4. Fogs

4.4.1. Definition

NOAA 1995: Fog is a collection of suspended water droplets or ice crystals near the Earth's surface that lead to a reduction of horizontal visibility below 1 km (smog: visibility below 10 km, airborne aerosols)

If the visibility is 1 - 10 km, then it is called mist (WMO, 1966)

The reduction of visibility occurs through a reduction in the brightness contrast between an object and its background by particle concentration and size-dependent scattering losses of the light propagating between the object and the observer (GAZZI et al., 1997; 2001) and through the blurring effect of forward scattering of light due to the presence of the droplets/crystals (BISSONETTE, 1992).

The differences between fog, mist and drizzle are vague.

4.4.2. Classification of fogs

Classification can be based on the following properties:

- physical properties (freezing/non-freezing fog)
- thermodynamic properties (mixed phase fog)
- dynamic properties (mixing and turbulence fogs)
- chemical composition of particles (smog, dry fog)
- orography of the surface (valley fog)
- meteorological features (frontal fog)

Mixed phase: the saturation temperature of a mixture can differ from those of its components. A fog that occurs as a front is passing by overhead is related to the mixing of nearly saturated warm and cold air masses.

Fog types are often classified by their formation mechanism. Even this classification is vague, however, because most often fog is the result of more than one process:

- radiation fog
- advection fog
- orographic fog (upslope fog)
- frontal fog or precipitation fog
- steam fog

4.4.3. Radiation fog

Radiation fog forms over non-frozen ground in the evening (= when the radiation balance becomes negative) in humid and almost clear weather with weak winds. If moisture is ample, the fog layer will grow higher all night as the upper parts cool continuously at a rate of roughly 1 °C an hour.

Favourable circumstances for the forming of the fog are low-lying regions (marshlands, river valleys) with strong evaporation during the day, or clear nights following rainy days.

The visibility in radiation fog rarely falls under 100 m and its thickness is only a few dozen meters. Dense and long-lasting fogs can occur in assosiation with anticyclones (especially in spring and autumn), when dust particles and condensation nuclei on the top of an inversion cool the air with their thermal radiation, generating a cloud layer that grows towards the ground.

The fog clears if there is sufficient insolation. The radiation heats not just the fog layer itself, but also the ground below it. The surface, in turn, heats the thin layer of air near it, thus generating a weak convective flow that mix this warm air with the lower parts of the fog. When this happens, the bottom of the fog thins and lets in more insolation, accelerating the dissipation of the fog, which proceeds from the ground up.

Fog can also dissipate due to wind. A strengthening wind mixes drier air with the fog at both the top and bottom of the fog layer. On the other hand, if an advection brings drier above the fog, the radiative cooling of the top of the fog layer can intensify, causing an opposite result.The temperature advection that accompanies wind can hasten the fog's dissipation. In a warm advection fog particles dissipate into water vapor. A cold advection near the top of the fog layer can also advance the dissipation by weakening the inversion that inhibits mixing.

The formation and dissipation of radiation fog are also affected by other clouds. The radiative cooling of the fog's upper parts during the night is strongest if there are no other clouds above the fog. A cloud layer coming over the fog (lower or middle clouds, or a thick layer of upper clouds) decreases the radiative cooling, thus also decreasing the formation of fog particles at the top of the fog. This can lead to a gradual dissipation of the fog if the weakened condensation can no longer keep pace with the forces working to dissipate the fog (of which the falling of fog particles, although slow, is always active).

Contrary to nighttime, clouds above a fog will slow the dissipation process during the day by reducing the amount of insolation that reaches the fog cloud and surface.

In late autumn the fog can either clear or not; it can also rise into stratus cloud and back again.

Radiation fog can occur where there is no wind, but weak winds contribute to its formation by lifting the turbulent mixing of air upwards after it has cooled near the surface, which creates a thicker layer of air favorable to the formation of fog. If the wind is completely still, the result can be no more than a thin layer of surface fog, dew or fog patches in cold, shallow depressions.

Note:

- radiation fog does not occur at sea
- radiation fog with a maximum thickness of 2 m is called surface fog.

4.4.4. Advection fog

Advection fog occurs when a warm and humid air mass cools as it passes over a cold surface.

Properties:

- dense
- relatively large
- often persists for a long time
- thickness usually around 300-500 m, but can be up to 1 km
- may rise upward and become an St cloud
- most common at sea, where the border of the fog is a vertical wall
- moderate winds, at least initially

In Finland, advection fog occurs particularly in spring and early summer in connection with flows from the land to the sea, when the warm air cools over cold water. On the other hand, fog also occurs often in autumn and early winter, when humid air is advected from ice-free seas over cold ground. In this situation the formation of fog is also aided by the continent's topography, which means that the thickest and longest-lasting fogs often form on hilltops. For example, the Rovaniemi airport, which is situated some 100 m above the Kemijoki valley,often experiences fog when humidity advecting from the Bay of Bothnia rises up the slope.

Fog dissipates if it is carried over a warm surface, or if there is sufficient insolation to dry it. In early summer the fog typically cannot advance farther than 10 km away from the shore.

4.4.5. Steam fog

This kind of fog occurs when cold air drifts over a warm sea and the temperature difference between them is large (c. 15 degrees or more).

The fog develops below a shallow inversion (in the absence of an inversion, the layer labilizes and the moisture rises higher). Situations like this occur near the coast; when the air travels above the sea for a long time, winds begin to undo the inversion. Sc clouds or shallow shower clouds are often formed over the sea. These clouds group together in lines and are sometimes arranged as roll clouds. Associated with sea smoke there is moderate wind or even gale.

Steam fog is common on the coasts of arctic regions in winter. In Finland sea smoke most often occurs from November through January, when the ground is frozen/covered in snow and the sea is free of ice. Non-frozen lakes steam in sub-zero temperatures, but usually there is only shallow fog (height below 2 m).

Larger-scale occurrences of steam fog are found on the fringes of cold ocean currents, such as in Newfoundland, northeastern United States, northern Pacific Ocean, the west coast of North America and the British Isles.

4.4.6. Precipitation fog

Precipitation fog forms when rain falls through air colder than itself and water vapour is evaporated.

A typical example of a situation like this is a large Nimbostratus cloud ahead of a warm front, which is why the fog is also called frontal fog.

4.4.7. Orographic fog

Orographic fog occurs when air flows up a slope, where

- the air is warm and humid
- the LCL is low
- the air is sufficiently stabile for there to be no convection

A fog is often a combination of orographic and advection fogs. The fog is dense and can persist for days. On the lee side of a hill the air descends and grows warmer, and and the fog dissipates at least partially.

Valley fog: there is a deep humid layer beneath an inversion. The inversion is a result of a long cooling period and descending motion of a persistent high pressure.

4.4.8. Dissipation of fog

Fogs dissipate through three mechanisms:

1. Temperature rises above the dew point

    - the surface and/or the fog absorb insolation
    - the fog moves over a warmer surface
    - the air heats adiabatically as it flows down a slope

2. Air humidity drops

    - water vapor condenses on a cold surface
    - snow falls through the fog layer and water vapor condenses on the surface of the crystals

3. The wind grows stronger

    - the fog rises up and becomes a stratus cloud
    - the fluxes of heat and humidity change, expediting the dissipation of radiation fog in particular

4.4.9. The effects of a snow cover

When air cools to a sub-zero temperature above snow, its water vapor condenses on the snow and the moisture flux is directed from the air into the snow. If the cooling is rapid and the humid layer thin, the moisture can condense on the snow and there will be no fog. However, if there is a lot of moisture, a snow cover will not prevent the formation of fog.

In some situations a snow cover can even advance the fog's formation and persistence:

- Snow does not warm up in sunlight, so when night falls the air cools faster over a snow cover than over non-frozen land, and the saturation point is also reached faster.
- Just like frozen ground, a snow cover also acts as an insulant, and there is no heat flux from the ground into the air.
- Thawing snow acts as a source of moisture as well.

4.4.10. Freezing fogs

Fog can occur in winter and although it is most likely to form and persist when air temperature is at or slightly below zero, it can also form in colder air. This kind of fog is called freezing fog, and it usually consist of supercooled water and ice crystals. Freezing fog is a relatively common phenomenon on the hills of northern and eastern Finland in winter. In the same conditions one can also find impressive rime formations and snow-loads on trees. Freezing drizzle can also occur at the same time with freezing fog.

Freezing fog and drizzle cause hazardous circumstances for traffic: ice forms on aircraft and ship structures and the windshields of cars, and road surfaces are covered with frost, causing skiddy conditions.

Fogs are rare in temperatures below -20 °C, but even then ice fogs can form. In this case the fog particles freeze or water vapor desublimates into ice crystals.

5. The basic properties of a wind field

Extra material: MetEd module 'Topics in Dynamic Meteorology: Thermal Wind'

5.1. Geostrophic wind and gradient flow

Let us take a look at horizontal air flows. We can assume that the following apply in the free atmosphere part of the troposphere in the mid-latitudes:

- isobars are straight and parallel
- there is no friction
- Rossby number < 0,1

Wind like this is geostrophic, and it is defined by the wind law

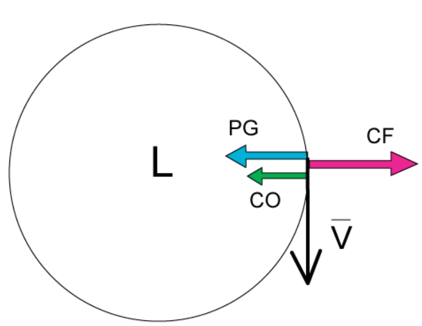
$$\overline{V}_g = \bar{k} \times \frac{1}{\rho f} \nabla_z p = \bar{k} \times \frac{g}{f} \nabla_p z$$

In other words, the height field determines the geostrophic wind.
In turn, a single height observation and the wind field determine the height field.

The influence of the Coriolis parameter makes it so that wind is about 1,5 times as strong in the Mediterranean as in the Arctic Ocean under similar pressure gradients.

And where does geostrophy not apply? When f is small or very large (tropics and polar regions), when the flow is not homogenous (wave disturbances, jet streams) or when it is influenced by friction (the boundary layer).

Gradient flow: a curved, geostrophic flow, where the pressure gradient force, Coriolis force and centrifugal force are in an equilibrium:



The equations of motion in a natural coordinate system (s,n):

$$\frac{dV}{dt} = -\frac{1}{\rho}\frac{\partial p}{\partial s} \qquad \text{s: along wind direction}$$

$$\frac{V^2}{R} + fV = -\frac{1}{\rho}\frac{\partial p}{\partial n} \qquad \text{s: along } 90^{o} \text{ to the left of wind direction}$$

R = the curvature radius of an air parcel's trajectory

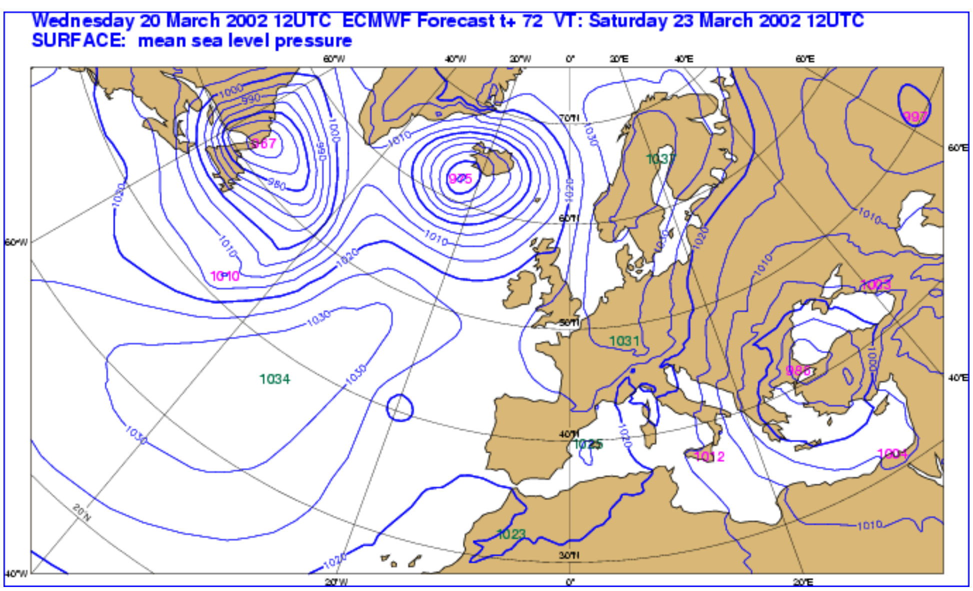
in a cyclonic flow R > 0
in an anticyclonic flow R < 0

curvature C = 1/R

The following apply to gradient flows:

- in a cyclonic flow V > V(g), pressure gradient is not limited
- in an anticyclonic flow V < V(g), pressure gradient is limited
- in a straight flow V = V(g)

Because of this the pressure fields of low pressures often have clear borders and tight pressure gradients, while the pressure fields of highs are more vague and their pressure gradients loose.



Ageostrophic wind = real wind – geostrophic wind

Ageostrophic wind is affected by four factors:

acceleration
advection
convection
friction

$$\overline{V}_{ag} = \frac{\overline{k}}{f} \times \left[\frac{\partial \overline{V}}{\partial p} + (\overline{V} \cdot \nabla_p \overline{V}) + \omega \frac{\partial \overline{V}}{\partial p} - \overline{F} \right]$$

I II III IV

I: isallobaric wind, towards greatest pressure drop

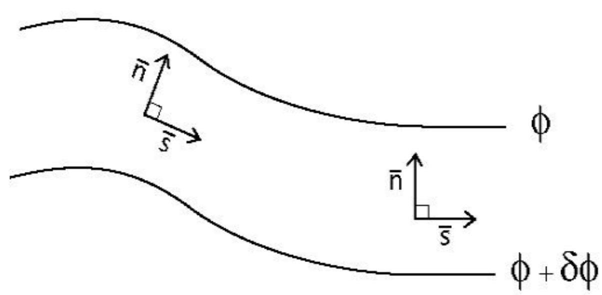
II: advective ageostrophic wind

III: convective ageostrophic wind (weak)

IV: friction

In the upper troposphere, the most significant of these terms is advection. Near the surface the friction term becomes significant.

in (n,s) - coordinates $\overline{V} \cdot \nabla_p \overline{V} = \bar{s}\, V \frac{\partial V}{\partial s} + \frac{V^2}{R} \bar{n}$



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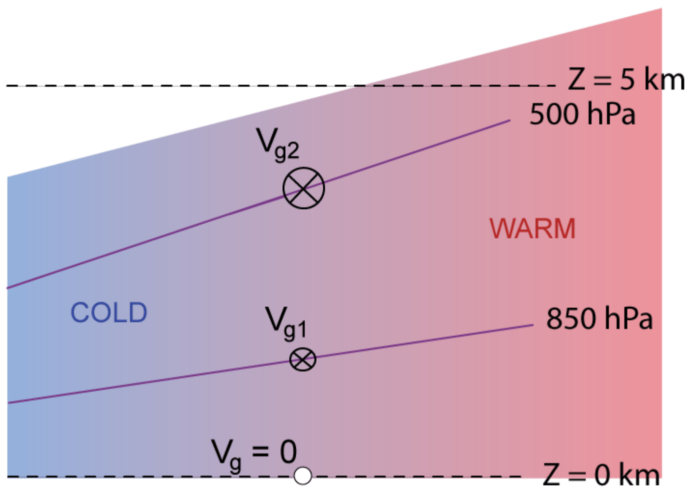
5.2. Thermal wind and temperature advection

Martin p. 89-93

Let us take a look at a layer of air where the temperature changes horizontally. Warm layers are thicker than cold ones, so the higher up the pressure surfaces are, the steeper their downward slope towards the cold air. The horizontal pressure gradient increases with altitude, and the geostrophic wind increases with it. The wind is represented by circles in the image below; the cross means that wind direction is into the page. Vg1 < Vg2.



So, there is a connection between the temperature gradient and the vertical shear of geostrophic wind. This connection is described by the thermal wind law, and next we will devise a mathematical representation for it. According to the hydrostatic equation

$$\partial p / \partial z = -\rho g ,$$

which can also be written as

$$\frac{g \partial z}{\partial p} = -\frac{1}{\rho} = -\frac{RT}{p} .$$

Because $\partial \phi = g \partial z$, the previous equation can be written as

$$\frac{\partial \phi}{\partial p} = -\frac{RT}{p}$$

, which is the isobaric form of the hydrostatic equation.

The geostrophic wind is $\overline{V}_g = (k/f) \times \nabla\phi$, and its vertical derivative

$$\frac{\partial \overline{V}_g}{\partial p} = \frac{k}{f} \times \nabla \frac{\partial\phi}{\partial p}$$

, which can be represented with temperature by using the isobaric form of the hydrostatic equation:

$$\frac{\partial \overline{V}_g}{\partial p} = \frac{k}{f} \times \nabla - \frac{RT}{p} = \left(\frac{-R}{fp}\right) k \times \nabla T$$

.

This vertical shear of the geostrophic wind is called a thermal wind:

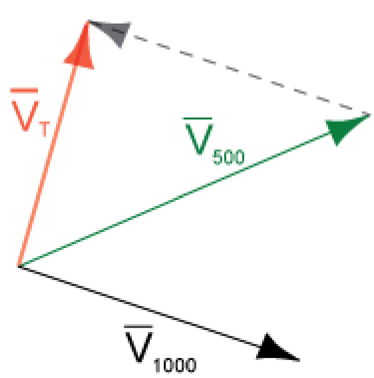
$$\overline{V}_T = \overline{V}_g - \overline{V}_o$$

In other words, a thermal wind is the difference between the geostrophic wind at an upper and a lower pressure surfaces. Its direction is parallel to the isotherms so that colder air remains on one's left when standing with one's back to the wind in the northern hemisphere.

Moving surface pressure anomalies travel in the direction of thermal winds (Martin p. 151-153).

If the layer is thin, its mean temperature can be assumed to be the mean value of the temperatures of its top and bottom surfaces.

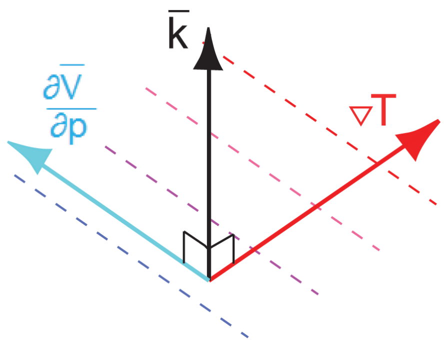
For example: $\mathbf{V}_T = V_{500} - \mathbf{V}_{1000}$:



Mathematically the thermal wind is defined as the cross product

$$\frac{\partial \overline{V}_g}{\partial p} = \left(\frac{-R}{fp} \right) k \times \nabla T$$

This means that, according to the cross product rule, the thermal wind is perpendicular to the vertical coordinate k and the temperature gradient dT- or, in other words, parallel to the isotherms. Due to the direction of the temperature gradient, the direction of the thermal wind is such that cold air is to the left and warm air to the right when one's back is to the wind in the northern hemisphere.



Because thermal wind strenghtens with altitude, and $\partial \overline{V}_g / \partial p < 0$, it can best be described as

$$V_T = -\partial \overline{V}_g / \partial p$$

Temperature advection TA

= scalar product of horisontal wind and temperature gradient, that gives the change in temperature caused by the wind:

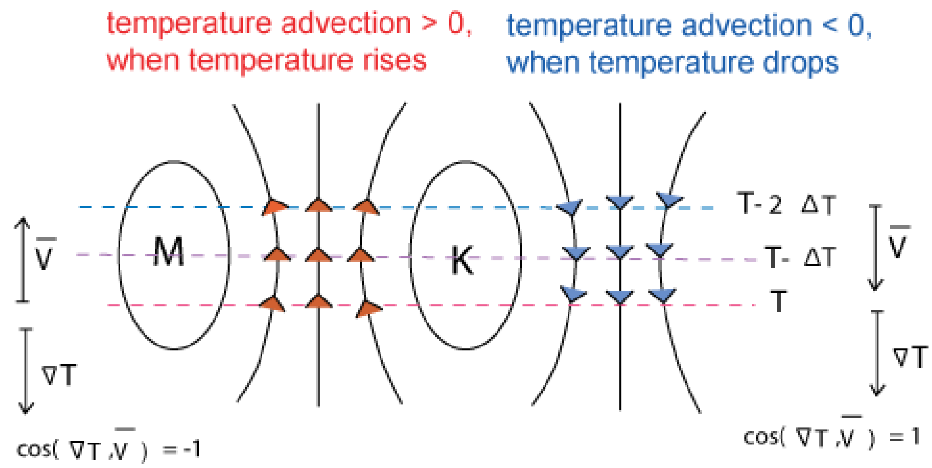
$$-(\overline{V} \cdot \nabla T) = -\left|\overline{V}\right| \cdot \left|\nabla T\right| \cos(\nabla T, \overline{V})$$

When TA > 0, there is warm advection and the air grows warmer.

When TA < 0, there is cold advection and the air cools.

The temperature gradient is a vector that points in the direction where the temperature increases most.

Note that temperature advection is therefore proportional to the temperature gradient, not temperature itself!

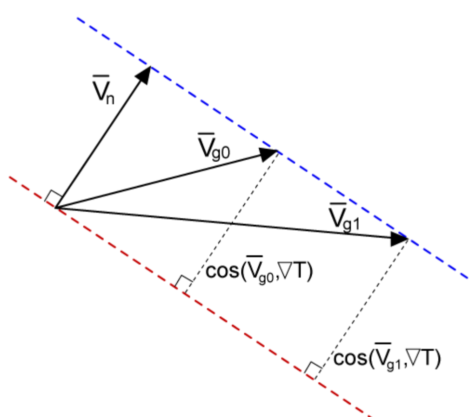


Temperature advection

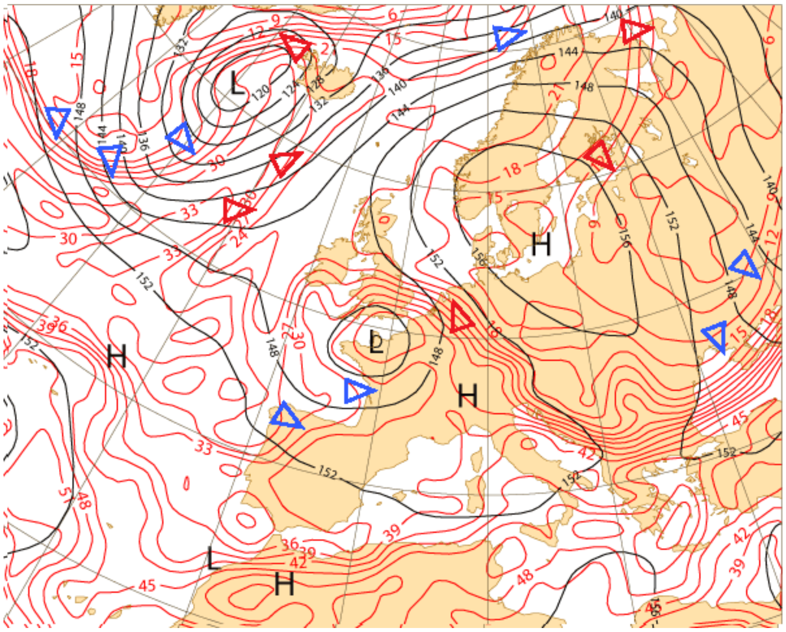
- largely determines temperature changes at 2 m
- regulates the development of upper waves
- influences vertical movements in the troposphere and, through them, precipitation
- regulates the development of baroclinic disturbances
- determines front type
- generates advection fog

Thickness advection describes phenomena on the synoptic scale better than temperature advection at a single pressure level.
If winds change linearly throughout a layer, thickness advection can be calculated from any wind in the layer in question.

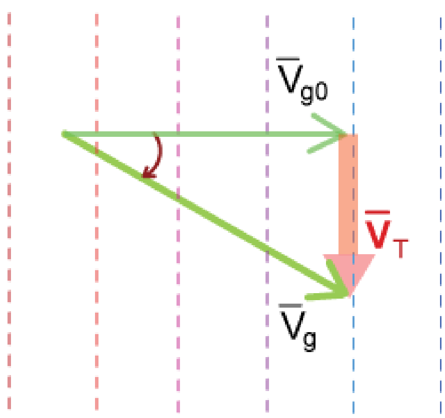
$-\overline{V}_g \cdot \nabla h = -\overline{V}_n |\nabla h|$ where h = the height of the layer,
because the wind perpendicular to the thickness gradient (V_n) is constant throughout the layer.



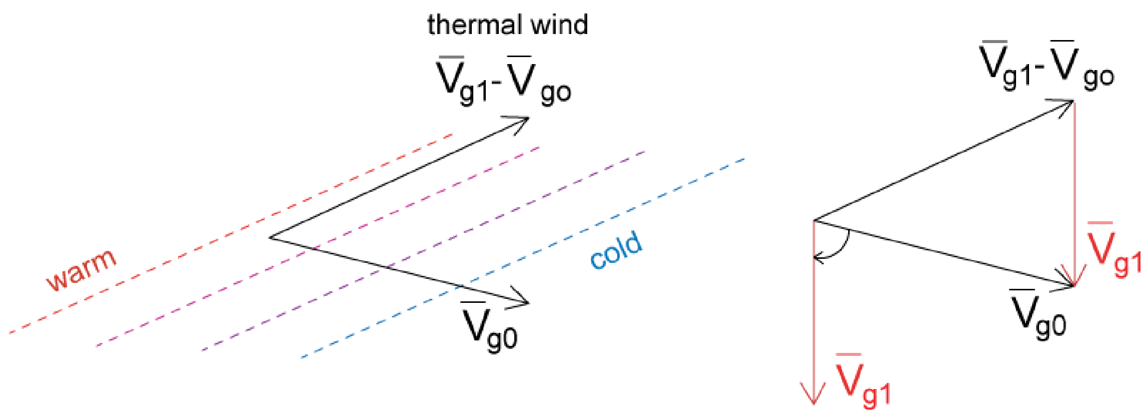
Temperature advection in a (1000 hPa altitude, 850 hPa thetae) field:



In warm advection the wind's direction is from warmer air towards cooler air, and vice versa in cold advection.



As the vector representation shows, in warm advection the wind veers with height. Correspondingly, in cold advection the wind turns counterclockwise:



Changes in tropospheric stability caused by warm advection (WA) and cold advection (CA):

Air becomes less stabile, when
- WA weakens with height
- CA strenghtens with height
- there is CA over WA
- there is CA at the upper levels

Air becomes more stabile, when
WA strenghtens with height
CA weakens with height
there is WA over CA
there is WA at the upper levels

5.3. Vorticity and its advection

Synoptics examines the vertical component of relative vorticity ζ.

In the troposphere vorticity generally equals angular velocity:

$$\eta = \zeta + f$$

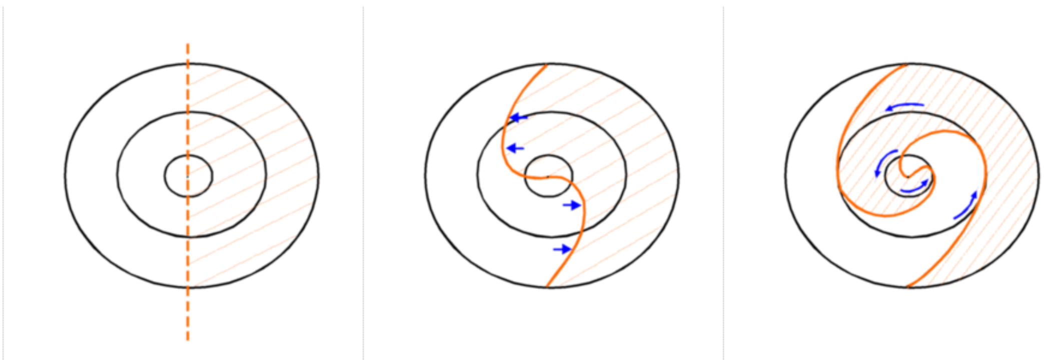
or, absolute vorticity = relative vorticity + the Coriolis parameter

Relative vorticity in a natural coordinate system (s,n):

$$\zeta = \frac{V}{R_s} - \frac{\partial V}{\partial n} \quad ,$$

where s = direction of the flow, n = direction perpendicular to the flow, and R(s) = the flow's radius of curvature.

The effect of cyclonic vorticity on cloudiness or a temperature field:



The components of relative vorticity:

curvature vorticity

$$\frac{V}{R_s}$$

shear vorticity

$$-\frac{\partial V}{\partial n}$$

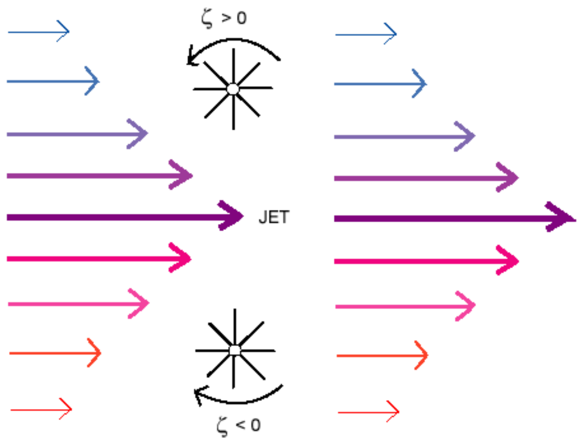
R_s = radius of curvature

Note that when vorticity ”decreases,” it becomes more anticyclonic. Its absolute value does not necessarily decrease.

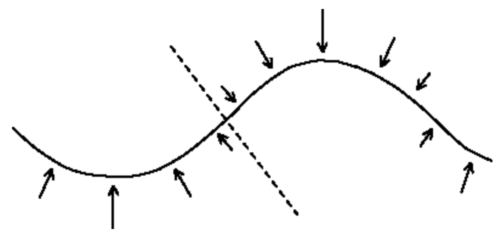
Shear: change in wind speed, curvature: change in wind direction

A practical example of shear vorticity: a paddle wheel placed inside a westerly jet stream rotates clockwise on the warm side of the jet, and counterclockwise on the cold side:

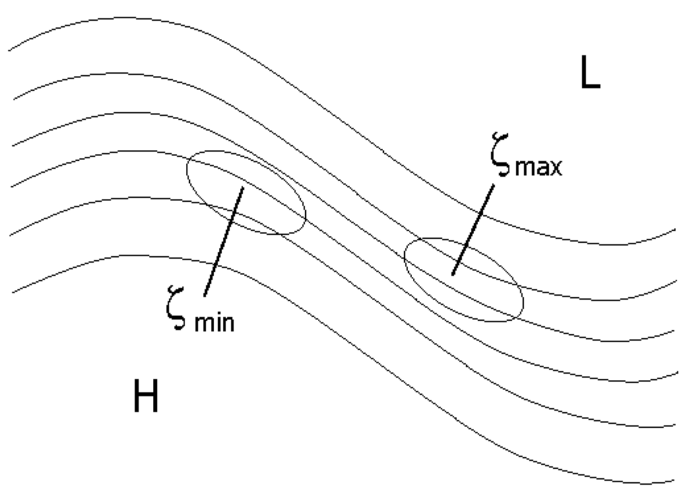
s-component:



n-component:



In a curved jet both components affect the total vorticity:



Shear vorticity maxima are connected with the entrance and exit regions of the jet cores while curvature vorticity maxima are connected with troughs and ridges.

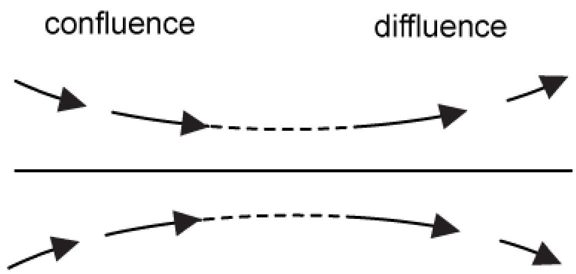
Vorticity advection

Horizontal advection of the vertical component of vorticity

- regulates movements of upper waves
- greatly influences vertical movements (and thus precipitation)
- is important for the initiation of wave disturbances

Positive vorticity advection, PVA, is often available in numerical models.

Confluence and diffluence in a flow:



Vorticity advection can be illustrated in a natural coordinate system by defining confluence curvature C(n) as follows:

$$VC_n = \frac{\partial V}{\partial s}$$

Confluence curvature is positive in a confluent flow and negative in a diffluent one.

In a natural coordinate system, absolute vorticity equals

$$\eta = VC_s - \frac{\partial V}{\partial n} + f$$

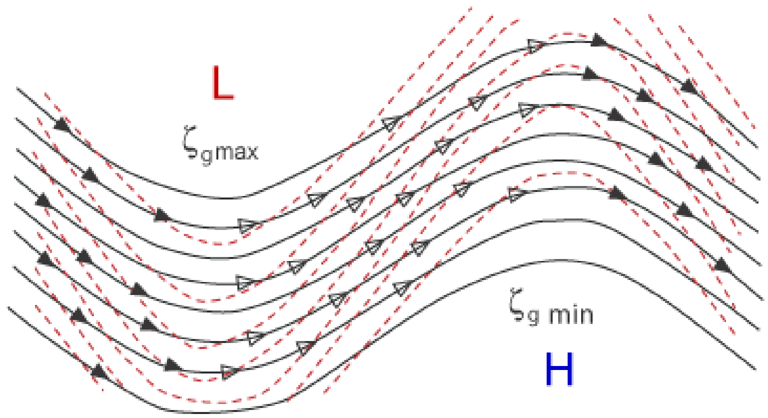
and the vorticity advection equals $A_\eta = -\overline{V} \cdot \nabla\eta$

Expressed with confluence curvature: $A_\eta = V^2(-\frac{\partial C_s}{\partial s} - C_s C_n)$

C(s) = 1/curvature radius.

Vorticity advection is proportionate to the square of wind velocity, which means the strongest vorticity advections are found in the jet streams of the upper troposphere.

Vorticity advection in an upper wave with no shear (with only curvature vorticity advection present):



The open arrows denote positive advection, and the black arrows denote negative advection.

$$A_\eta = V^2\left(-\frac{\partial C_s}{\partial s} - C_s C_n\right)$$

where C(s) > 0 in cyclonic flows and C(n) > 0 in confluent flows.

Regions of positive vorticity advection:

- anticyclonic confluence
- cyclonic diffluence
- cusp line ahead of a trough

Regions of negative vorticity advection

- anticyclonic diffluence
- cyclonic confluence
- cusp line ahead of a ridge

6. Jet streams

6.1. Properties of jet streams

A jet stream is a local wind maximum. The streams are thousands of kilometers long and hundreds of kilometers wide (though branches can be shorter).

Jet streak is the maximum of the jet stream. Jet streaks travel in the direction of the stream at a velocity far smaller than that of the wind maximum. The velocity of air particles increases at the entrance of the jet streak and decreases at the exit.

Numerical models often present isotachs in intervals of 10 m/s, starting at 30 m/s.

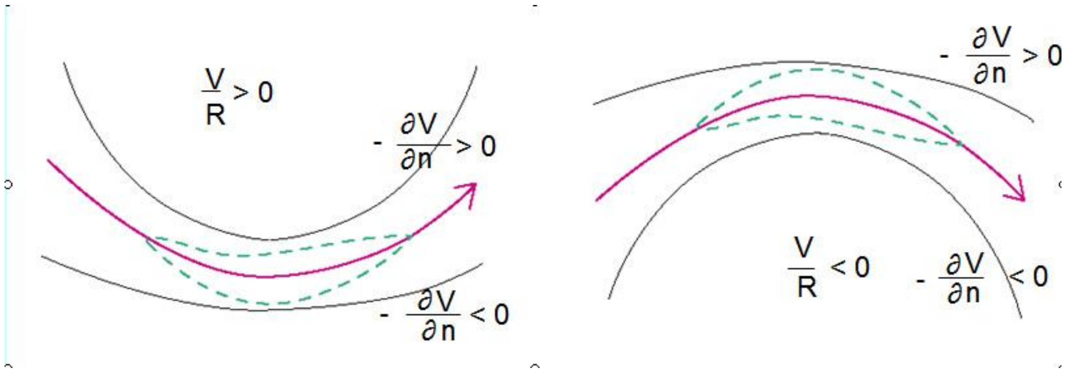
6.2. Jet streams and vorticity

Vorticity is proportionate to the square of wind speed, so the highest values of vorticity are found in upper troposphere jets:

- high shear vorticity in entrance and exit regions
- curvature vorticity throughout the region

Sign of vorticity

- jet at the bottom of a trough: curvature vorticity positive
- jet at the top of a ridge: curvature vorticity negative
- shear vorticity positive on the cold (north) side of the jet, negative on the warm (south) side



The sign of total vorticity therefore depends on one's position regarding the jet stream, how curved the stream is and how strong it is.

Wind on the cold and warm sides of a jet stream

Absolute vorticity has to be positive in the troposphere for the air staying stable:

$$\eta = \frac{V}{R_s} - \frac{\partial V}{\partial n} + f \geq 0$$

So, in a straight flow:

$$- \frac{\partial V}{\partial n} \geq -f$$

Shear vorticity is always positive in a cyclonic flow, but the equation above limits the distance between isotachs in an anticyclonic flow:

$$\Delta n \geq \frac{\overline{\Delta V}}{f}$$

6.3. Jet streams and divergence

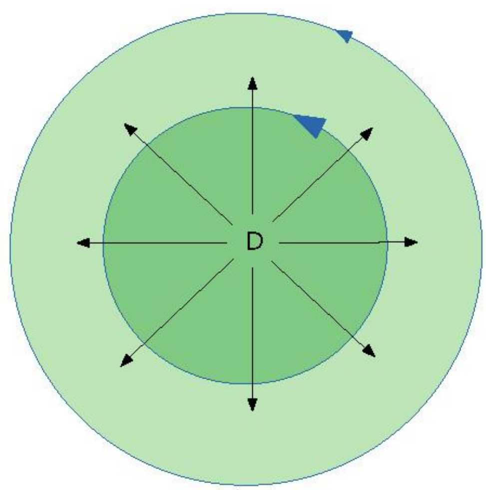
Divergence and vorticity (Martin p. 132-133)

If there is divergence in a flow element, its area increases with time.

vorticity = circulation/area

So, If the area increases, vorticity decreases (becomes more anticyclonic).

In the following schematic the blue arrows in the figure represent vorticity and the green areas represent divergence:



Consequently, if there is convergence in the flow, vorticity increases (becomes more cyclonic).

Cf. the spiral of a figure skaters: when skaters bring their arms close to their body (convergence), angular velocity (vorticity) increases.

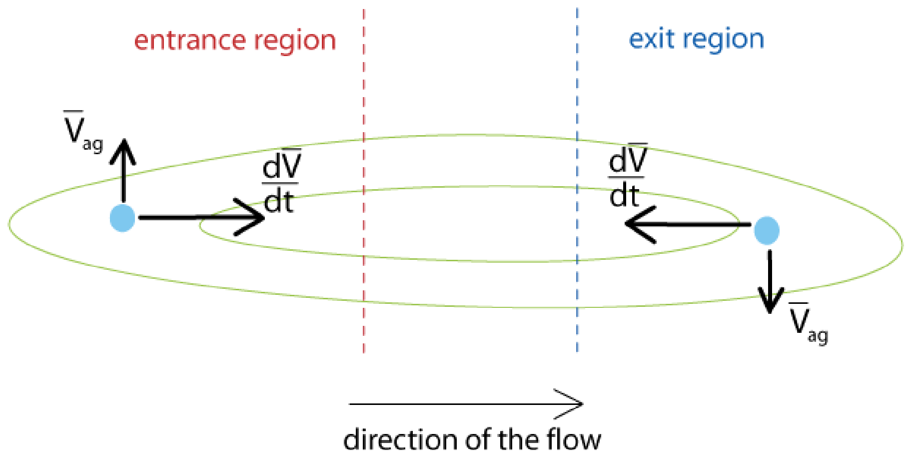
6.4. Ageostrophic wind in a jet stream (Martin p. 148-149)

Geostrophy does not apply in jet streams, which involve strong variation in wind speed.

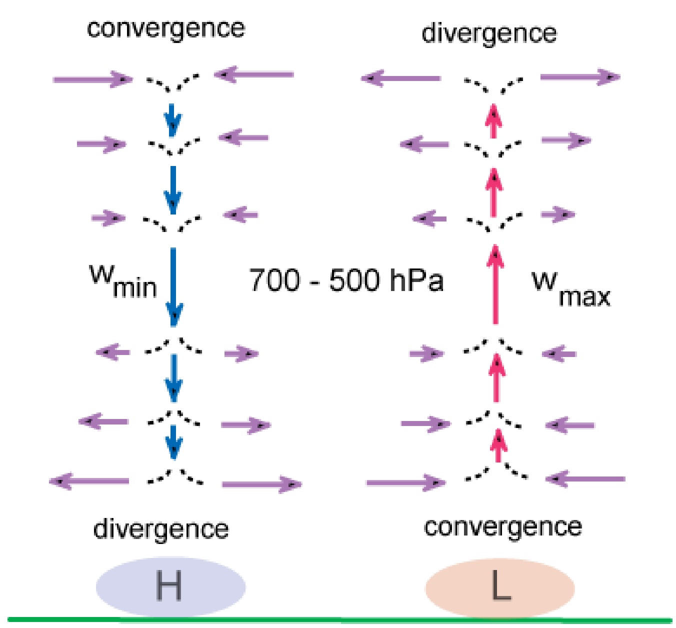
In a straight jet stream the ageostrophic wind equals

$$\bar{V}_{ag} = \frac{\bar{k}}{f} \times \frac{d\bar{V}}{dt}$$

Ageostrophic wind causes the real wind direction to deviate from the direction of the geostrophic wind:



The connection between divergence and vertical movements:

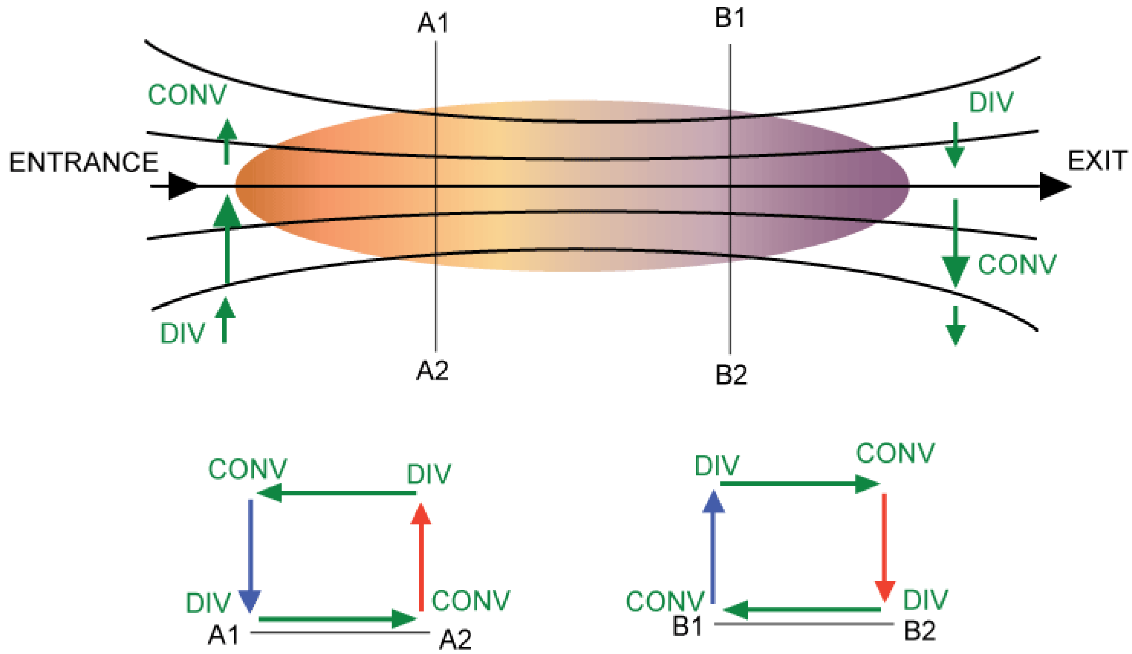


The practical consequence of this is that there is ascending motion and often precipitation in the leading edge of an upper trough, while in the leading edge of an upper ridge there is descending motion and often fair weather.

Precipitation is least likely in Finland when the wind is northwesterly and there is a ridge of high pressure over the Scandinavian Mountains; hence the Finnish folk saying, "A northwesterly wind is the sky's broom."

In a straight jet stream there is divergence and convergence in velocity.

The ageostrophic wind component causes divergence in the right entrance and left exit of the jet stream, which produces rising motion under these regions. There may be heavy precipitation in these areas.

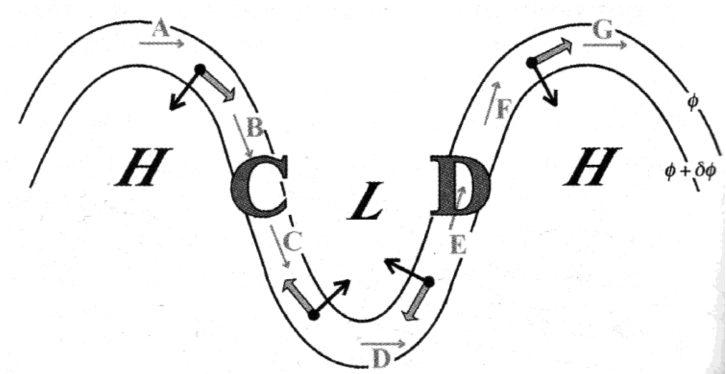


(Nielsen: Quasigeostrophic baroclinic development)

In the entrance region the air parcels accelerate. The acceleration maximum is on the jet axis. The acceleration is proportional to the ageostrophic wind, which blows from south to north and increases towards the jet axis. This leads to divergence in the right entrance region (A2), and the warm air rises below it.

In the exit region the air parcels decelerate and the ageostrophic wind blows from north to south. There is divergence in the left exit region (B1) and the cold air rises below it.

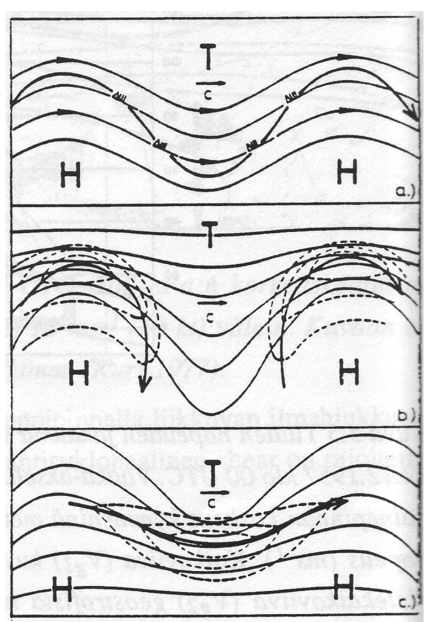
In a curved flow with identical wind speed throughout its volume, wind acceleration is determined solely by changes in wind direction:



(Martin, Figure 6.2.)

In point A the wind is westerly, in point B northwesterly. There has to be southwestward-directed acceleration between them for the wind to turn. The wind does not change direction between points B and C, and so, no acceleration there.

The ageostrophic wind causes the deviation of real wind direction from geostrophic wind direction, which means the jet stream axis of a curved jet stream crosses the isohypses of the height field:



(M. Kurz: Synoptic Meteorology 1989)

In other words, the shear in a jet stream generates divergence and convergence. On the other hand, the more cyclonically curved the flow is, the higher its curvature vorticity $V \cdot C(s)$.

Geostrophic wind is unbounded in a cyclonic gradient flow and bounded in an anticyclonic one. Typically

$$V_g(\text{trough}) \approx 3 \cdot V_g(\text{ridge})$$

Which means jet streams are at their strongest in troughs.

In summary:

A strong, cyclonically curved jet stream:

$\Longleftrightarrow$ high positive vorticity advection

$\Longleftrightarrow$ convergence and divergence

$\Longleftrightarrow$ vertical movements

$\Longleftrightarrow$ precipitation

According to the thermal and geostrophic wind laws:

jet stream in the upper troposphere (200-400 hPa)

$\Longleftrightarrow$ high pressure gradient in the upper troposphere (300–500 hPa)

$\Longleftrightarrow$ high temperature gradient in the mid-troposphere (500-850 hPa)

$\Longleftrightarrow$ high thickness gradient in the lower troposphere (850-1000) hPa

7. Synoptic scale vertical motions

7.1. Divergence

Martin p. 81

Divergence is present when more air leaves a location than comes to it. Convergence is present when more air comes to a location than leaves it.

When an air parcel is caught in a flow with divergence, it stretches and its surface area grows. In convergence it shrinks and its surface area diminishes.

In a pressure coordinate system, we can determine divergence from the continuity equation with the hydrostatic assumption:

$$D = \nabla \cdot \bar{v} = -\frac{\partial \omega}{\partial p} ,$$

In other words, horizontal divergence on an isobaric surface is directly proportional to the vertical motion.

Note 1: The sign of $\omega < 0$ in ascending motion, $\omega > 0$ in descending motion.

Note 2: Small values of divergence do not have any noticeable effect on the weather. A typical value of divergence in weather disturbances is 10(exp)-5 1/s.

Ascending motion in troposphere: air has to rise when there is convergence in the lower part, because it cannot go anywhere else. And, correspondingly, below the tropopause air arrives from below, so it has to spread horizontally, since it can go no higher.

$$D = -\frac{\partial \omega}{\partial p}$$

At the surface and in the tropopause $\omega = 0$. The pressure surface where vertical motion achieves its maximum value lies somewhere between the two. At this level the derivative of divergence with regard to pressure equals 0, which means that divergence on that surface also equals 0. On the other hand, the derivative of a function equals 0 at the inflection point, so the divergence changes sign on this surface.

Therefore, when going from the surface to the tropopause, the sign of the divergence will change at least once.

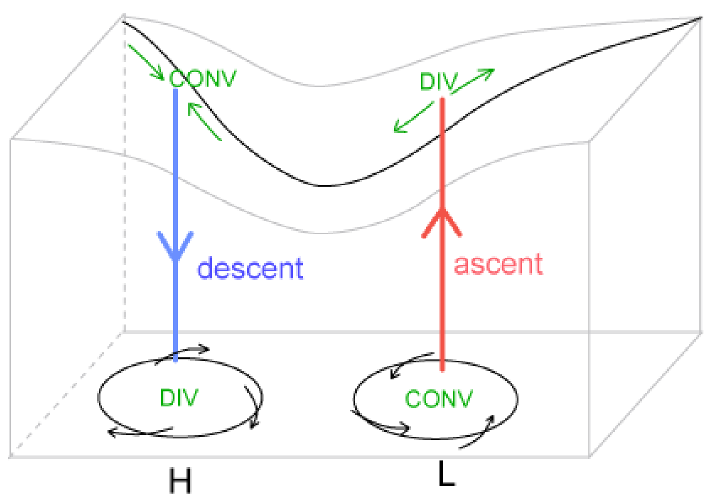
By integrating the above equation from surface p(0) onto the first non-divergent surface p(n), we get

$$\bar{D}\,(p_o - p_n) = -(\omega_n - \omega_o) = \omega_o - \omega_n\,,$$

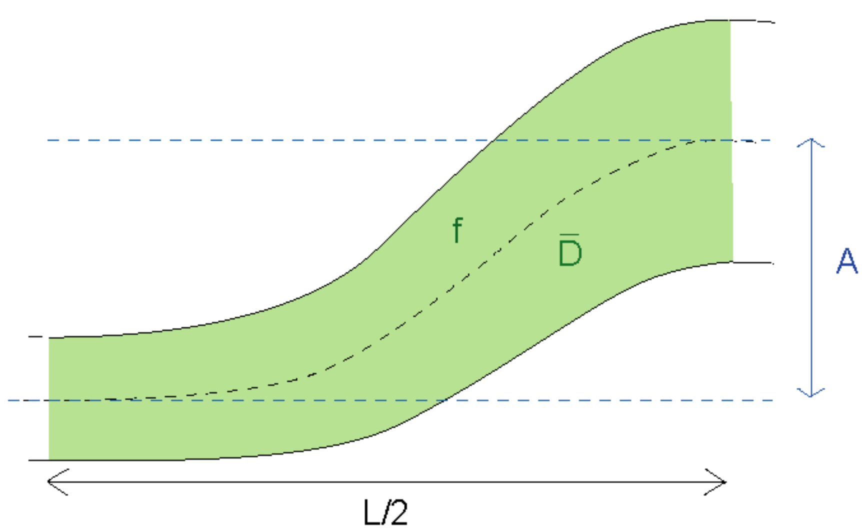
where $\bar{D}$ is the mean divergence of the layer p(o) - p(n).

$$\omega_o = 0 \ \text{ and } \ p_n < p_o \quad \Longrightarrow \quad D = -\frac{\omega_n}{p_n - p_o}$$

which means $\bar{D}$ and ω(n) are of the same sign.



Palmen, Newton 1969:



$$\overline{D} \approx \pm \frac{16\,\pi^2 A}{f\,L^3}\; V(V - V_n)$$

Here V is the gradient wind and V(n) wind on a non-divergent surface.

The divergence of the gradient wind is strongest when the wave amplitude is at its highest, the wavelength short, and wind speed high and different from the wind of the non-divergent surface.

V - V(n) is greatest at the height of the wind maximum, so that is also where divergence and convergence are at their peak.

However, V - V(n) is also a thermal wind, so divergence in the upper troposphere and vertical movements in the mid-troposphere depend on the flow's baroclinity.

7.2. Sutcliffe Development Theorem

Martin p. 157 -160

Sutcliffe's theory from 1939 was one of the first operationally practicable theories in flow dynamics. Sutcliffe later refined the theory by adding the geostrophic assumption.

In the derivation of the theorem the following assumptions has been made:

- horizontal and vertical winds are geostrophic, and there is no friction
- the vertical advection of vorticity is negligible
- the relative vorticity of the divergence term is negligible
- adiabatic factors are discounted
- deformation terms (the products of derivatives) are neglected
- the layer is represented by its average geostrophic and thermal winds

With the aforementioned assumptions, the theorem can be written as

$$f_o(\nabla \cdot \overline{V} - \nabla \cdot \overline{V_o}) = -\overline{V_T} \cdot \nabla(\zeta_{g_o} + \zeta_o + f),$$

where o = the lower altitude, and V_T = the thermal wind

According to this equation, the difference in wind divergence between two pressure (or geopotential) surfaces equals the vorticity advection brought by the layer's thermal wind. If the difference is positive, the advection of cyclonic vorticity is present.

On the other hand, as mentioned earlier, divergence being larger high up than down results in rising motion.

In other words:

- ascending motion when the thermal brings cyclonic vorticity advection
- descending motion when the thermal brings anticyclonic vorticity advection

In reality, vertical motions are much more complicated than this, but the theorem presents a sufficiently accurate approximation of large-scale vertical motions to be used operationally.

7.3. Omega equation

Martin p. 162 - 166

The quasigeostrophic vorticity equation
The thermodynamic energy equation

The omega equation for vertical motions on the synoptic scale:

$$\underbrace{\sigma\left(\nabla^2 + \frac{f_o^2}{\sigma}\frac{\partial^2}{\partial p^2}\right)\omega}_{A} = \underbrace{f_o \frac{\partial}{\partial p}\left[\overline{V}_g \cdot \nabla(\zeta_g + f)\right]}_{B} + \underbrace{\nabla^2\left[\overline{V}_g \cdot \nabla\left(-\frac{\partial \phi}{\partial p}\right)\right]}_{C}$$

Ascending motion: $\omega < 0$, descending motion: $\omega > 0$, σ = stability index.

$f_o = 2\Omega \sin\phi$ f(0) is often assumed constant.

The equation contains no time derivatives, so it is a diagnostic equation for ω, presented by instantaneous geopotential. It provides a measure of vertical motions with no need for accurate wind observations.

Term A:

$$\sigma(\nabla^2 + \frac{f_o^2}{\sigma}\frac{\partial^2}{\partial p^2})\omega$$

Vertical motions can be described with a sine function. Generally, the derivative of sine Dsinx = cosx, and Dcos = -sinx, so

$$\frac{\partial\ \omega}{\partial p^2} \propto -\ \omega$$

and further:

$\nabla^2\omega$'s min/max is also ω's min/max

So when the right side of the equation is positive (negative), there is ascending (descending) motion.

Term B:

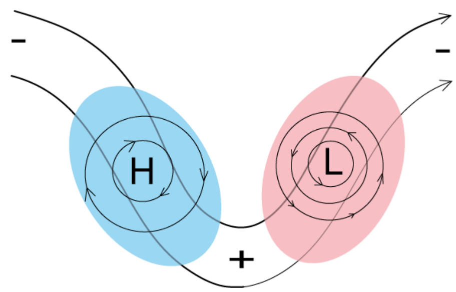
$$f_o\frac{\partial}{\partial p}[\bar{V}_g \cdot \nabla(\zeta_g + f)]$$

= the vertical derivative of geostrophic vorticity advection

The term is positive when cyclonic vorticity advection increases with height, i.e. it generates ascending motion.

If the term is negative, cyclonic vorticity advection decreases (and anticyclonic grows) with height, and the motion is descending.

Relative vorticity advection is more important in synoptic scale vertical motions than planetary vorticity advection.



The image above shows an upper wave above a high and a low at the surface. Vorticity advection is weak at the surface because the vortex is limited to a restricted area and the circulation is nearly closed. The upper wave contains strong vorticity (+ cyclonic, - anticyclonic), so positive vorticity increases with height over the surface low and decreases with height over the surface high (i.e. negative vorticity increases with height). Ascending motion (red area) develops above the surface low and descending motion (blue area) develops above the surface high.

Term C

$$\nabla^2[\overline{V}_g \cdot \nabla(-\frac{\partial\phi}{\partial p})] = -\nabla^2 \underbrace{[-\overline{V}_g \cdot \nabla(-\frac{\partial\phi}{\partial p})]}_{\text{horizontal temperature advection}}$$

The term is positive when there is a local maximum of warm advection, and negative when a local maximum of cold advection is present.

Note that vertical motion is not determined by the temperature advection itself, nor even the change thereof, but rather the change in temperature advection's change (the inflection point of temperature advection). If the temperature advection gradient is zero, meaning the temperature advection is constant everywhere, or if the advection changes uniformly everywhere, no vertical motion will develop.

A summary of the omega equation

Ascending motion is caused by

- positive vorticity advection that increases with height
- a maximum of warm advection

Descending motion is caused by

- negative vorticity advection that increases with height
- a maximum of cold advection

Ascending motion is strongest when stability is weak.

Vorticity advection is weak at the surface, so usually it is enough to look at vorticity advection at 500 hPa. Temperature advection, on the other hand, is strongest at 850-700 hPa.

Vorticity advection and temperature advection can also counteract each other.

Terms B and C can be combined by assuming f to be constant, applying the definition of the Jacobian determinant to the advection and assuming the deformation of thermal wind to be small, which yields us this approximation:

$$\sigma \left(\nabla^2 + \frac{f_o^2}{\sigma}\frac{\partial^2}{\partial p^2}\right)\omega \approx 2\left[f_o \frac{\partial V_g^2}{\partial p} \cdot \nabla(\zeta + f)\right]$$

Kurz 1977: if s is defined to be parallel to the isotherms and n the direction of -∇T in a natural coordinate system, the equation can be approximated (w = -ω):

$$w \approx \frac{\partial \zeta_g}{\partial s} \cdot \frac{\partial T}{\partial n}$$

However, $\partial T / \partial n < 0$, so:

Ascending motion is present in the vicinity of a large temperature gradient when the thermal wind blows from a region of higher vorticity to a region of lower vorticity. Accordingly, therethere is descending motion when the thermal wind blows from a low-vorticity region to a high-vorticity one.

7.4. Forced vertical motion

orography: the effects of the altitude of the surface (mountains, valleys, etc.)
topography: the effects of the properties of the surface (land, sea, snow, etc.)

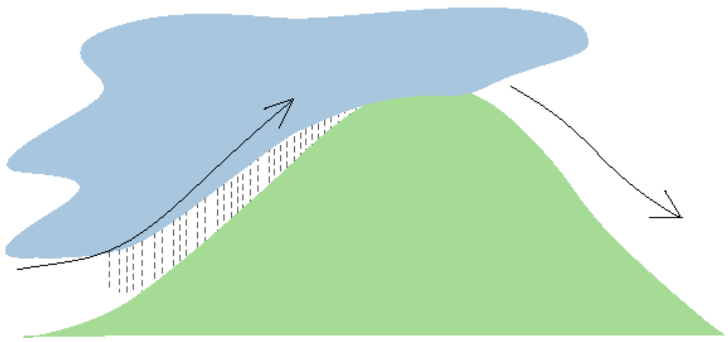
Orographic vertical motion:

When a sufficiently humid air flow rises up a slope, the air cools and the water vapor condenses into clouds and precipitation.The lee side experiences dry adiabatic descending motion, precipitation ceases and clouds dissipate

Orographic vertical motions are common in the Scandinavian Mountains.

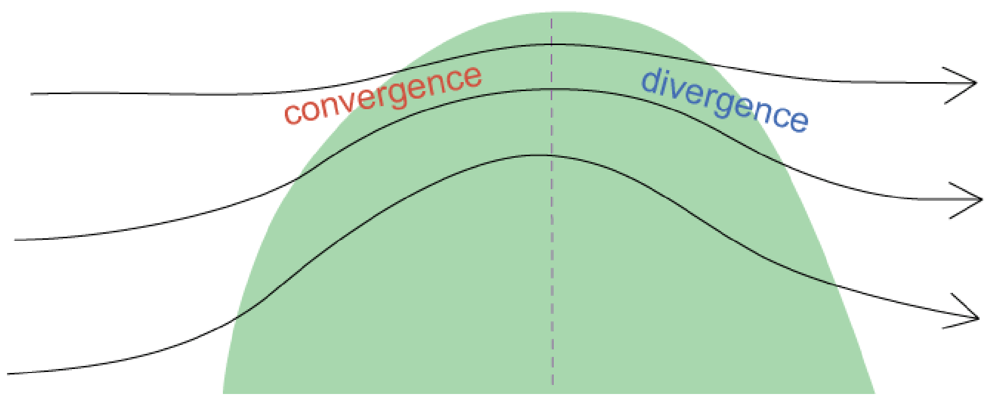
The record high temperatures in winter in western Lapland (as high as +10 °C) are the result of warm Föhn winds coming down the slopes.

When a synoptic scale area of ascending motion crosses a mountain range, the orography strengthens precipitation on the windward slope and weakens it on the lee side. Showers and thunder may develop in potentially unstable conditions.



Mountain range seen from behind

Upon meeting an end of a mountain range, an air flow with try to avoid it by going round the edge (‘shoulder effect’), which generates convergence on the windward side and divergence on the lee side. This can be seen in precipitation statistics (Bergeron, southern Norway 1949).



End of mountain range seen from above

Other vertical motions caused by the surface:

- drainage flows from cold mountain slopes (glaciers, snowy mountains)
- coastal convergence (land – sea)
- frictional convergence (field – forest)
- effects are modest in size, but visible in precipitation statistics

7.5. The quasi-geostrophic height tendency equation Martin p. 247-249

The tendency equation deals with changes in geopotential with time.

Baroclinic zone = a zone with a higher temperature gradient than its environment
Baroclinic zones are related to upper waves and jet streams.

The development of baroclinic waves can be examined with the help of temperature advection and vorticity advection. The essential applications are:

- the development of surface highs and lows
- vertical motions
- the transformations of air masses

The tendency equation uses the so-called quasi-geostrophic model:

- horizontal geostrophy
- vertical hydrostaticity

$\Rightarrow$ thermal wind law

Other approximations:

- frictional forces are small
- vertical advection is weak
- tilting term is small in the vorticity equation
- z is small compared to f
- diabatic heating is weak
- the scale of horizontal motion is much smaller than the Earth's radius (a ”beta-plane approximation”, where f is constant)

These assumptions are the same as in Sutcliffe's theory.

On a closer look, diabatic heat transfer (the release of latent heat) influences the development and intensity of cyclones significantly. We will return to this in chapter 12.

The tendency equation:

The rate of geopotential change, χ , is

$$\chi \equiv \frac{\partial \Phi}{\partial t} \approx g \frac{\partial z}{\partial t}$$

It can be described with the quasi-geostrophic height tendency equation:

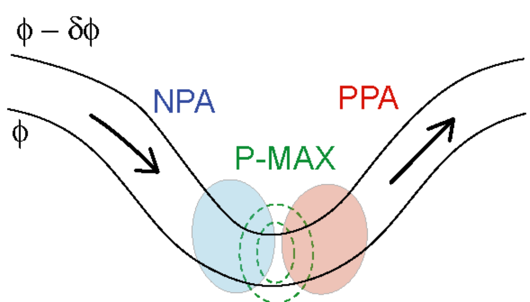
$$\underbrace{\left(\nabla^2 + \frac{f_o^2}{\sigma}\frac{\partial^2}{\partial p^2}\right)\chi}_{\substack{\text{A} \\ \text{change of height}}} = \underbrace{-f_o \bar{v}_g \cdot \nabla \left(\frac{1}{f_o}\nabla^2 \Phi + f\right)}_{\substack{\text{B} \\ \text{vorticity advection}}} + \underbrace{\frac{f_o^2}{\sigma}\frac{\partial}{\partial p}\left(-\bar{v}_g \cdot \nabla \frac{\partial \Phi}{\partial p}\right)}_{\substack{\text{C} \\ \text{differential} \\ \text{temperature advection}}}$$

In a statically stable atmosphere $\sigma > 0$.

$f_o = 2\Omega \sin\phi_o$

Term B describes the effects of geostrophic vorticity advection on local changes in the height field.

An upper trough with the associated geostrophic wind, curvature vorticity maximum as well as negative and positive vorticity advection:



Vorticity advection is zero at the bottom of a trough and at the top of a ridge, so geopotential does not change there. This means that vorticity advection neither strengthens nor weakens the wave, but rather merely causes it to move forward (or backward, as in some cases).
The term can also be written as

$$-f_o \bar{v}_g \cdot \nabla \left(\frac{1}{f_o}\nabla^2 \Phi + f\right) = \underbrace{-\bar{v}_g \cdot \nabla\left(\frac{g}{f_o}\nabla^2 z\right)}_{B1} \underbrace{- v_g \frac{\partial f}{\partial y}}_{B2}$$

B1: advection of relative vorticity
B2: advection of planetary vorticity

In waves on the synoptic scale, B1 > B2,
In waves on the planetary scale, B1 < B2.

The physical interpretation:

Relative vorticity advection tends to move the upper waves along the flow, and planetary vorticity advection tends to move them against the flow.

The real movement depends on the scale of the disturbance:

- synoptic waves (L< 5000 km) move with the flow
- planetary waves (L > 10 000 km) move against the flow

The aforementioned only applies to westerly flows. In other cases:

- B1 and B2 have the same sign in easterly flows
- B2 is negative in southerly flows (troughs weaken and ridges intensify)
- B2 is positive in northerly flows (troughs intensify and ridges weaken)

However, the waves are relatively short in these cases, so B2 has little significance.

The effect of vorticity advection according to wave length:

- synoptic waves travel with the flow
- planetary waves travel against the flow
- waves in between are stationary

Term C is the so-called differential temperature (or thickness) advection

$$\frac{\partial}{\partial p}\left(-\overline{v}_g \cdot \nabla \frac{\partial \Phi}{\partial p}\right) \approx -\frac{\partial}{\partial p}\left(-\overline{v}_g \cdot \nabla \frac{1}{\rho}\right) \approx -\frac{\partial}{\partial p}\left(-\overline{v}_g \cdot \nabla T\right)$$

The geopotential tendency is proportionate to the change in temperature advection with height, or to the differential temperature advection.

Wind in the upper troposphere is often roughly parallel to the thermal wind (i.e. isohypses and isotherms are parallel). The temperature advection is strongest in the lower troposphere, weakening with height.

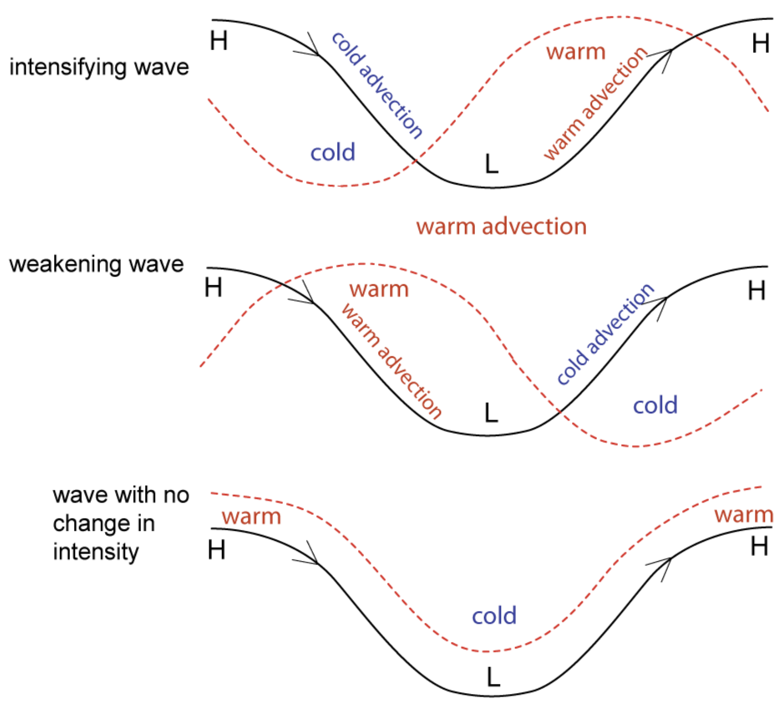
Estimation of geopotential tendency at 500 hPa with the differential temperature advection between 850 and 300 hPa (WA=warm advection, CA=cold advection):

300 hPa	0	0	CA	WA	WA	CA	0	WA	CA
850 hPa	WA	CA	WA	CA	0	0	0	WA	CA
χ(500 hPa)	> 0	< 0	>> 0	<< 0	< 0	> 0	0	$\approx$ 0	$\approx$ 0

Strong temperature advections often cause either warm or cold advection throughout the troposphere.

If temperature advection decreases with height, differential temperature advection is negative and geopotential increases, and if temperature advection increases with height, differential temperature advection is positive and geopotential decreases.

What the term C means in practice: if there is upwards weakening warm advection in an upper ridge, and upwards weakening cold advection in the accompanying upper trough, the upper wave intensifies. As the advections in a cyclone development are strongest in the lower troposphere, weakening with height, it can be stated briefly that if there is warm advection in a ridge and cold advection in a trough, the wave (and cyclogenesis) intensifies.



7.8. Vertical motion and clouds

Frequently and on average:

- strong ascending motion in the mid-troposphere: Ns, thick As or Ac
- strong descending motion: clear skies
- descending motion high up: St, Sc, Cu or low Cb
- ascending motion high up: As, Ac or Ci

Synoptic scale numerically derived parameters (temperature advection, vorticity advection, omega, etc.) cannot describe:

- thin layered clouds formed by mixing
- convection
- mesoscale vertical motions related to fronts (precipitation bands, etc.)
- some orographic and topographic clouds

Of all clouds, the Sc, St and Ac types are particularly notable, since they are usually not related to synoptic scale vertical motions (‘(for example: subsidence inversion, spreading of cumulus) and may cause incorrect forecasts especially with regard to temperature (it is cooler in summer and warmer in winter under a cloud layer than without it).

In Finland, one source of nasty surprises in forecasting are cold northeasterly flows in summer, which may carry a persistent low cloud layer that results in cool temperatures. In general, continental air flows may contain thin, low-lying inversions that escape numerical models.

8. The development of high and low pressures

A pressure anomaly is an area of higher or lower pressure than its surroundings. In height and pressure fields it is seen as a closed curve.

Low pressures contain cyclonic, and high pressures anticyclonic vorticity.

Anomalies that travel with the upper waves are called cyclones and anticyclones. They are connected with temperature gradients in the lower troposphere. A strong thermal wind is necessary for anomalies to develop, which makes them baroclinic disturbances.

Stationary highs and lows:

- thermal highs and lows
- blocking highs
- cut-off lows in the mid- and upper troposphere

8.1 The formation and disappearance of a surface low

Terminology:

lows deepen and highs weaken when $\partial p/\partial t < 0$

lows weaken and highs strengthen when $\partial p/\partial t > 0$

lows intensify when $\partial \zeta/\partial t > 0$

highs intensify when $\partial \zeta/\partial t < 0$

The quasi-geostrophic vorticity equation (for example: Holton):

$$\frac{\partial \zeta}{\partial t} = \underbrace{-u\frac{\partial \zeta}{\partial x} - v\frac{\partial \zeta}{\partial y} - \beta v}_{-\vec{V}\cdot\nabla(\zeta+f)} \underbrace{-(f+\zeta)\nabla\cdot\vec{V}}_{+(\zeta+f)\frac{\partial \omega}{\partial p}}$$

or

$$\frac{\partial \zeta}{\partial t} = -\overline{V} \cdot \nabla(\zeta + f) - (f + \zeta)\nabla \cdot \overline{V}$$

This means vorticity changes due to the influence of advection and divergence.

Winds are weak near the surface, resulting in a small advection term, and vorticity develops almost solely due to divergence.

Solving for divergence from the continuity equation shows that the distribution of vertical motions affects the development of vorticity.

According to the omega equation, vertical motion is determined by vertical changes in vorticity advection and changes in temperature advection.

Sutcliffe and Forsdyke, 1950: relative vorticity on a lower surface, $\zeta_{(go)}$, and on a higher surface, $\zeta_{(g)}$, are connected in the following way:

$$\underset{A}{\frac{\partial \zeta_{g0}}{\partial t}} = \underset{B}{\frac{\partial \zeta_{g}}{\partial t}} - \underset{C}{\frac{g}{f}\nabla^2 \frac{\partial h}{\partial t}} + \underset{D}{\left(\frac{\partial \zeta}{\partial t}\right)_F},$$

where F refers to friction. Term D is only significant when the combined effect of the other terms is small and a clear center of low or high pressure already exists at the surface.

This equation allows us to estimate the development of high and low pressures. The frictional term is relatively small in the development stage, so high and low pressures develop when

- strong positive vorticity (term B) comes over the pressure anomaly
  or
- the temperature inside the air column changes considerably in a limited area (term C)

Depressions form in front of upper troughs. They will deepen if

- the upper trough deepens,
- the upper flow accelerates, or
- there is warm advection ahead of the upper trough

High pressure forms in front of upper ridges. It will intensify if

- the upper ridge strengthens,
- the upper flow accelerates, or
- there is cold advection ahead of the upper ridge

Note: the vorticity equation does not say whether the anomaly at the surface is one of high or low pressure! For example, the environment of a rapidly filling low pressure center can contain descending motion and divergence.

The intensity of the development of surface pressure anomalies can be measured with

- divergence in the upper troposphere
- ascending motion in the mid-troposphere

Regarding the intensity of pressure anomalies:

- vorticity is unbounded in low pressures; values as high as $\zeta = 5f$ have been observed
- in high pressures $-\zeta < f$, because absolute vorticity in the troposphere > 0

The disappearance of a surface low

A low begins to fill when

- vorticity advection and changes in temperature advection counteract each other
- vorticity advection and changes in temperature advection grow weaker and the frictional term starts having an effect

Friction will typically exhaust a surface vortex in a couple of days. An upper vortex will usually persist a little longer.

The ascending motion caused by friction often generates a lot of St or Sc clouds in the vicinity of filling lows for as long as the flow remains cyclonic. Fog can also develop in this way.

8.2. The vertical structure of highs and lows

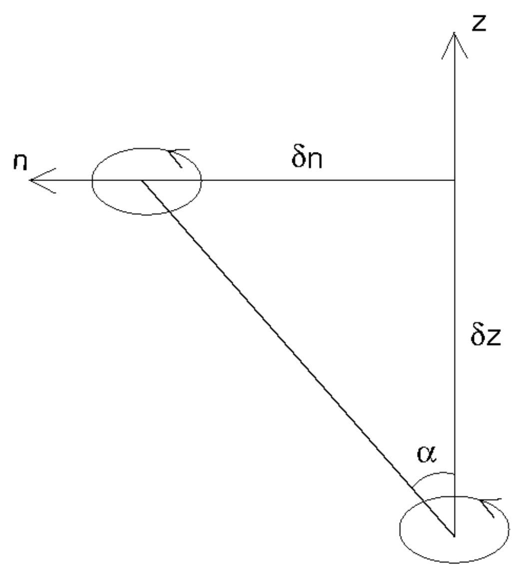
Surface troughs and ridges tend to develop near the inflection points of upper waves, where the vorticity advection is strong

troughs and ridges tilt upwards and against the flow in the formative stage.

The vertical axial tilt of pressure anomalies:

Kurz 1977: the axial tilt of a low pressure can be determined from changes in temperature and height:

$$\tan\alpha = \frac{1}{T}\left(-\frac{\partial T}{\partial n_p}\right)\left(\frac{\partial^2 z}{\partial n^2}\right)^{-1}$$



$\alpha > 0$

in a low pressure center $\nabla^2 z > 0$, so $-\partial T/\partial n > 0$

in a high pressure center $\nabla^2 z < 0$, so $-\partial T/\partial n < 0$

$\partial T/\partial n < 0$ in low pressure, so the axis tilts in the direction of where the air is coldest, and in high pressures toward where it is warmest.

The greater the horizontal temperature gradient above the surface low, the wider the angle of inclination.

Furthermore, the equation shows that the axis is upright when

- there is no horizontal temperature gradient, or
- surface vorticity is at the extreme point of the temperature field

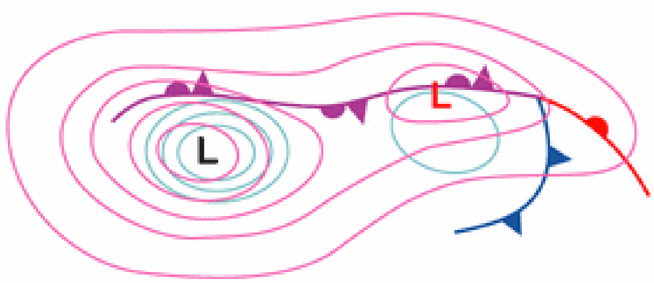
When the horizontal temperature gradient disappears, the low will no longer deepen. At that point the surface and upper lows lie on top of one another. However, what has been described up to this point has only been the prediction of the quasi-geostrophic theory; in practice, a cyclone's development can continue even after that, for example with the formation of a secondary low.

Secondary low

In certain conditions a secondary low will deepen within an occluded front: shear and curvature vorticity are both strong, the jet core is at least 50 m/s and it crosses the occlusion point.

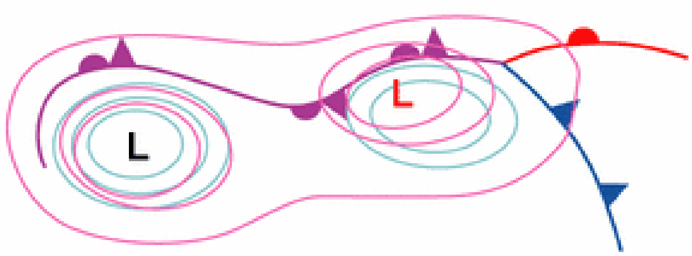
Schematics of the development:

1) a secondary low begins to deepen near the occlusion point, when the original low is already far from it

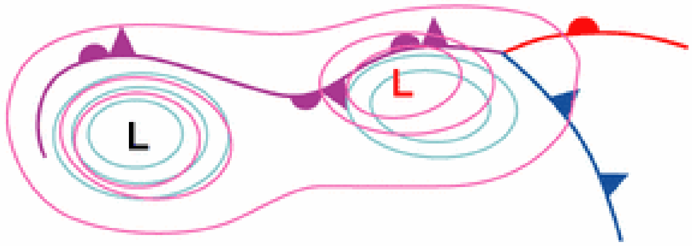


(pink: surface pressure, cyan: upper pressure)

2) the original low fills and the secondary one deepens, often becoming deeper than the original one



3) both lows fill



Vertical changes in vorticity

On a constant pressure surface, the hydrostatic equation and the equation of state give us

$$\frac{\partial}{\partial z}(\nabla^2 z) = \frac{1}{T}(\nabla^2 T)_p$$

In low pressure $\nabla^2 z > 0$. If it is colder than its environment, which is to say, $\nabla^2 T > 0$, the intensity of the low increases with height.

Cold low pressures intensify and warm ones weaken as height increases.

Cold low pressures are typical in the mid- and high latitudes. In the mature stages of baroclinic disturbances, high temperature gradients are located at the edges of upper troughs, and the upper trough is colder than its environment.

Beneath a cold upper low there can be a much weaker surface low, trough or even a weak high pressure. Lows like this are called cut-off lows or cold drops. Shower clouds and thunders often develop in them. Cut-off lows are studied in chapter 9.

9. Different types of high and low pressures

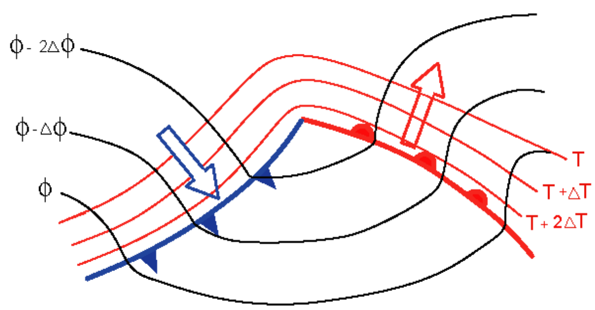
9.1.1. The Norwegian cyclone model

The Norwegian cyclone model, developed in the early 20th century, describes the development of a cyclone moving along a westerly flow. Analyzing fronts with this method remains a quick and usable way of interpreting weather conditions. The quasi-geostrophic theory deals with the same topic, but it refers to cyclones as disturbances in the baroclinic zone, or waves.

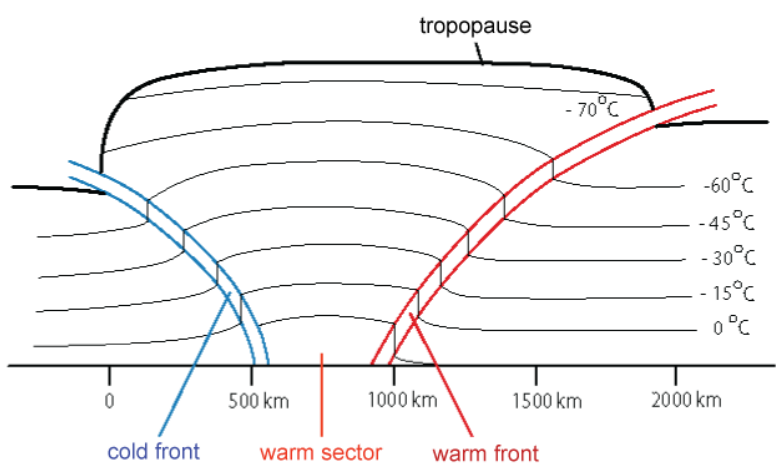
Waves in the baroclinic zone are interpreted as fronts:

- a zone with warm advection comprises a warm front
- a zone with cold advection comprises a cold front
- between the warm and cold fronts there lies a warm sector

The best place to search surface fronts with numerical models is the warm edge of either 850 hPa temperature gradient or potential temperature gradient:



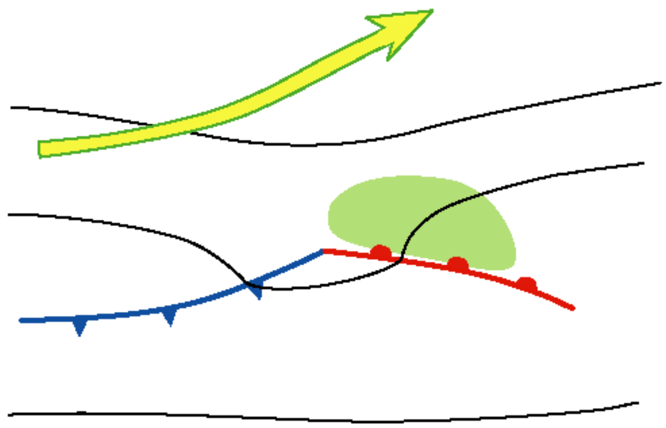
Fronts incline towards cold air with height. A vertical cross section of a wave:



A schematic depiction of cyclone development according to the Norwegian model:

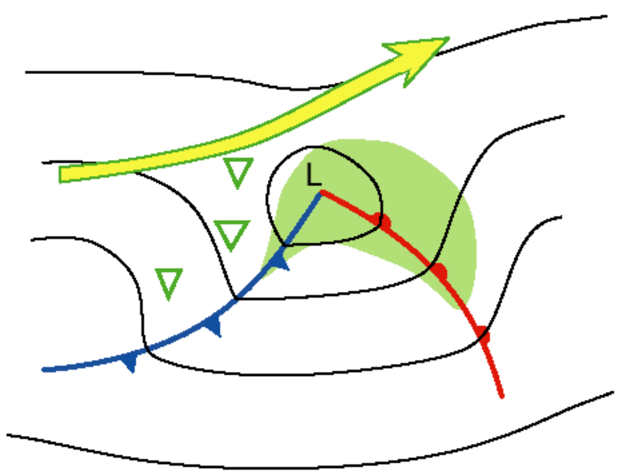
1. A wave forms

On the warm side of a polar jet, a trough in the surface pressure field and a wave in the temperature field begin to form.



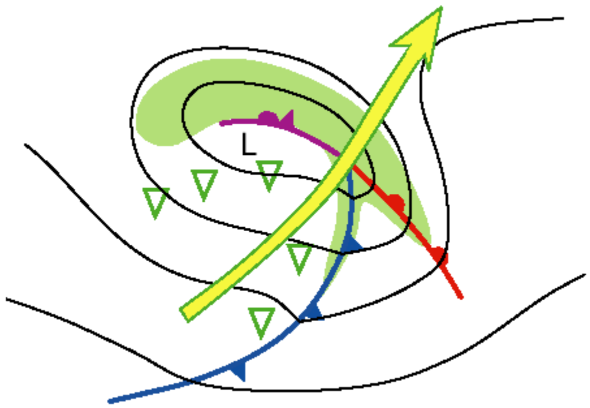
2. The wave intensifies

A low pressure develops at the surface. Ascending motion generates clouds and precipitation. The upper trough lying upstream of the surface low deepens.



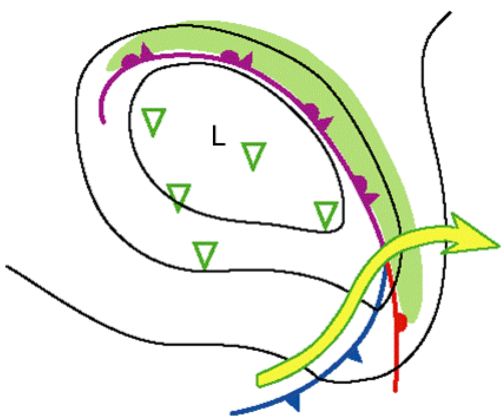
3. The wave occludes

Cold fronts move faster than warm ones. When the two merge and become an occluded front, the warm sector fades away starting from the bottom. Now there is also a low high up, and its axis of the low rises upright. The pressure stops falling and the jet stream passes over the occlusion point.



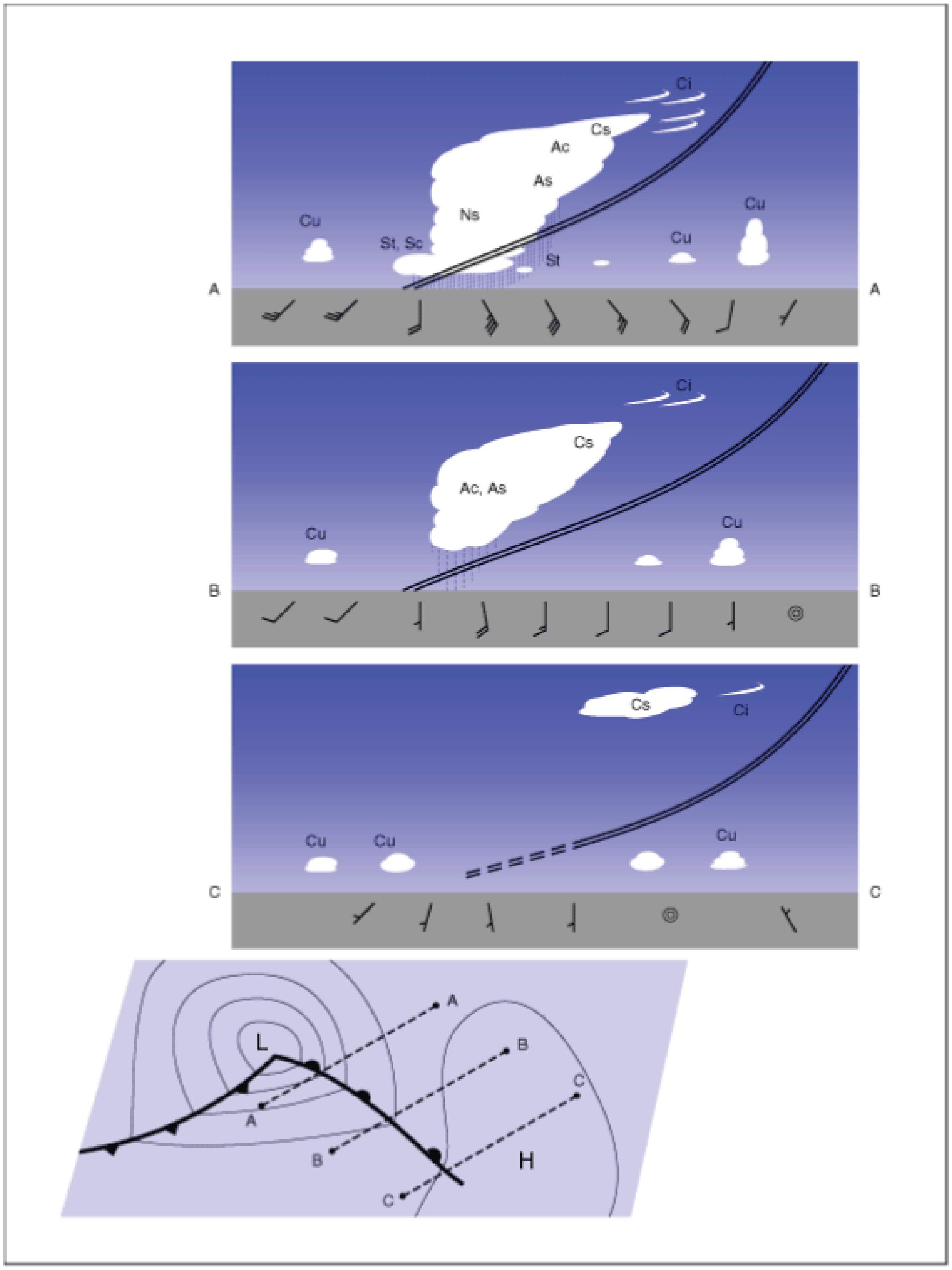
4. The disturbance disappears

Pressure rises and synoptic scale vertical motions cease. What remains is a weakening occluded front.

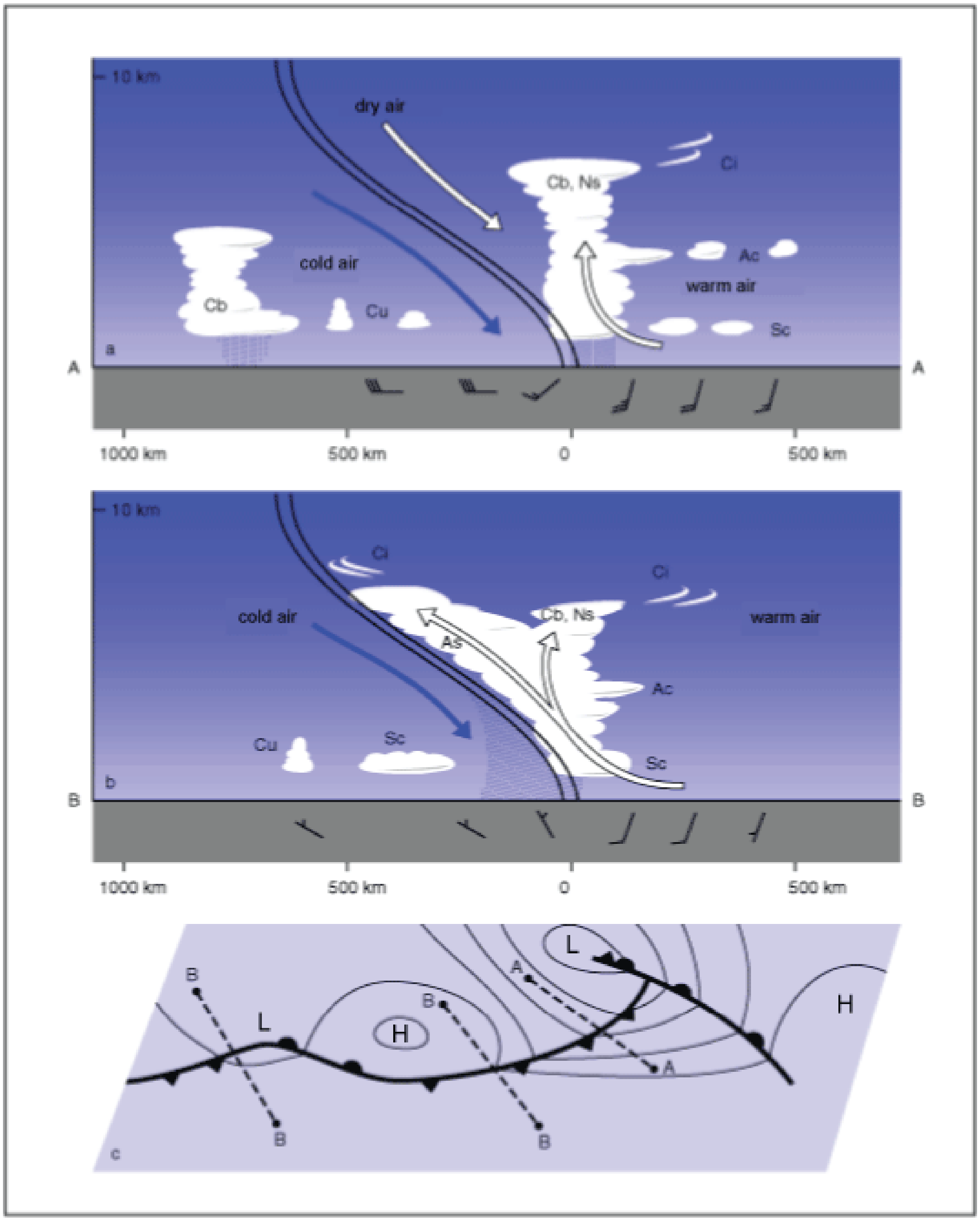


Karttunen et al: *Ilmakehä, sää ja ilmasto*:

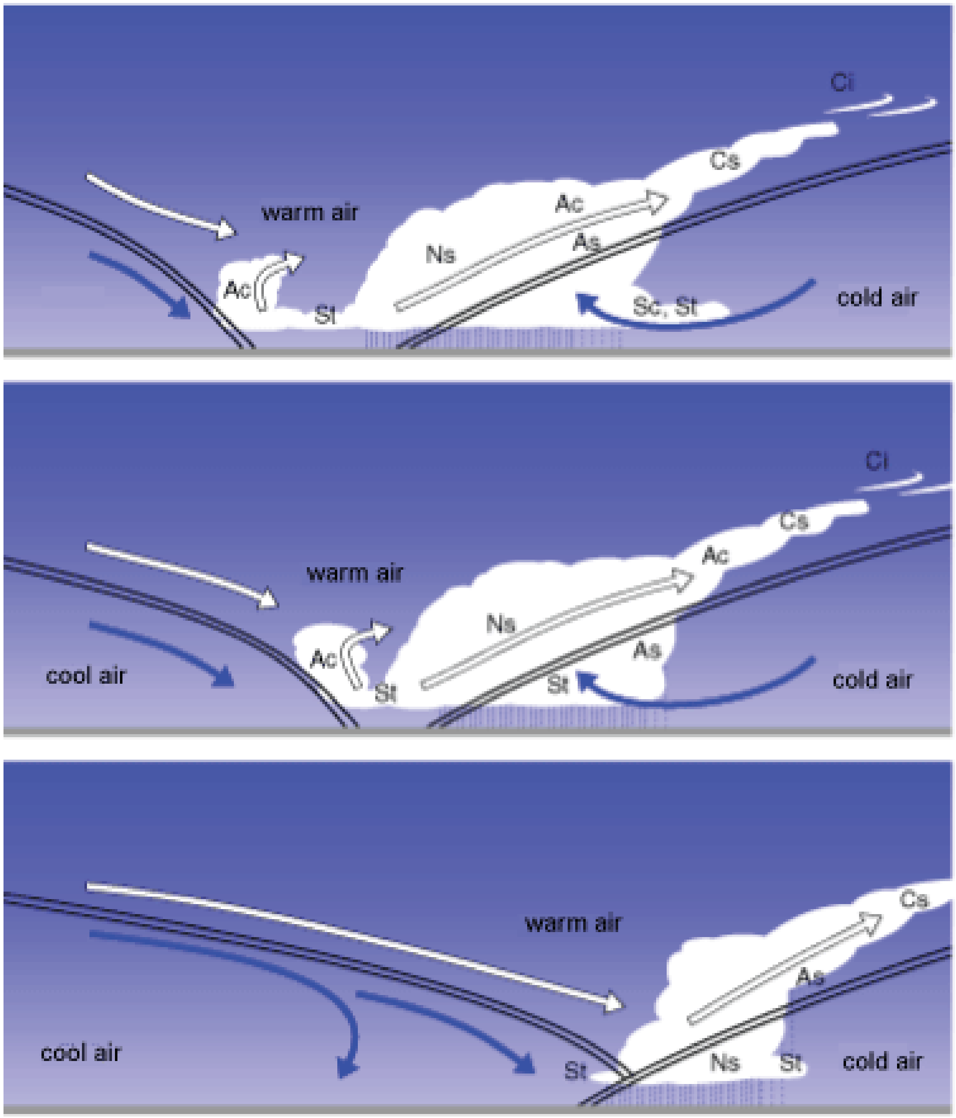
Clouds, winds and precipitation of a warm front in different distances from the low center:



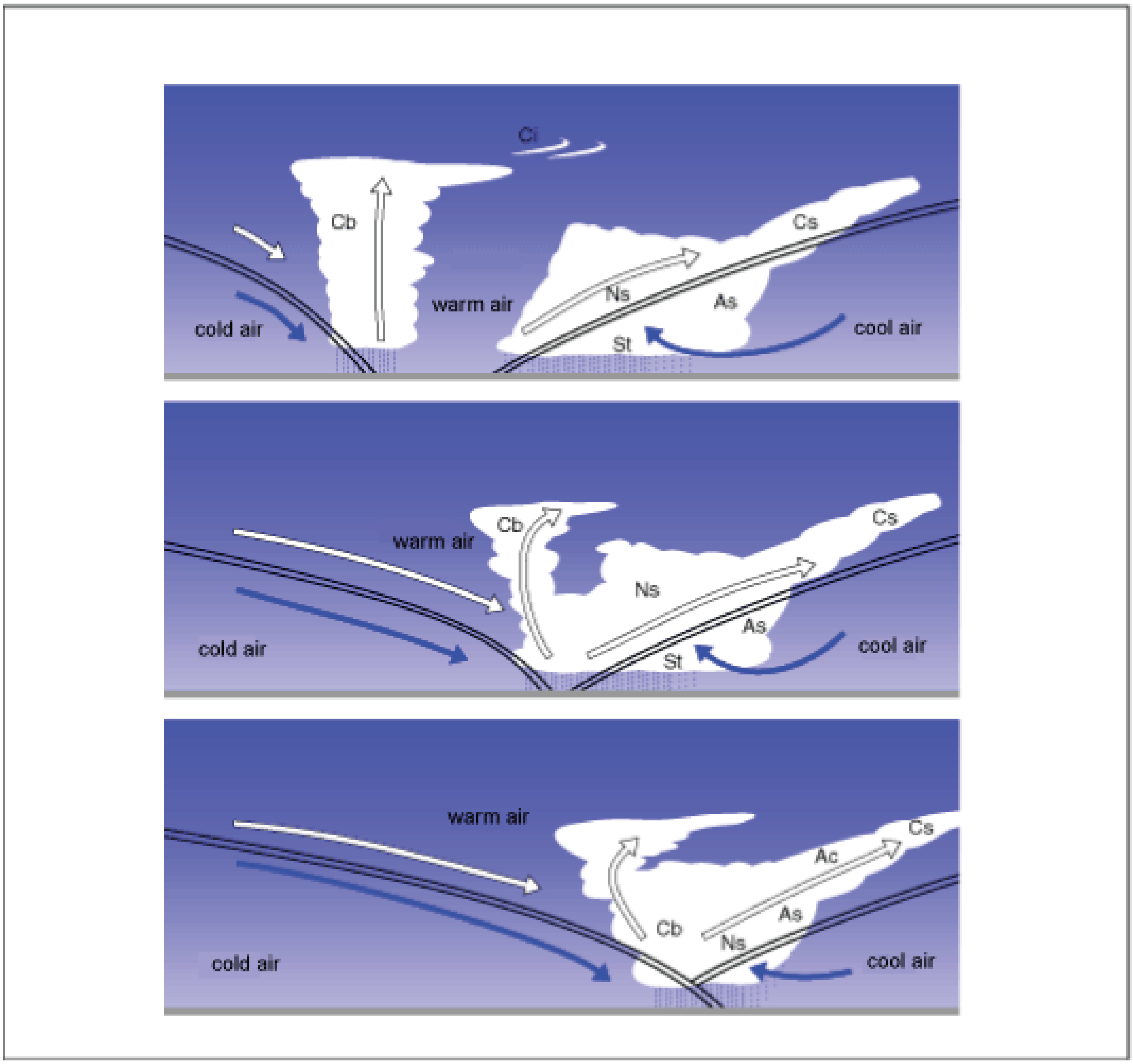
Clouds, winds and precipitation of a cold front in different distances from the low center:



Clouds, winds and precipitation of an warm occluded front:



Clouds, winds and precipitation of an cold occluded front:



Cyclones form often families: a new low deepens within the cold front of an occluded cyclone. This way cyclones follow each other in the westerly flow, until it weakens or becomes meridional.

A cyclone will, on average, travel 1000-8000 km with the 500 hPa flow during its development. The movement starts out fast and slows down towards the end.

Cyclones usually develop in the western Atlantic, reach their maximum intensity in the British Isles and western Europe, and die around Scandinavia or eastern Europe.

A cyclone's life span is most often 3-7 days.

Note:

Not all young wave disturbances develop into cyclones.

The Norwegian model does not work properly outside westerly flows. For example, lows that arrive in Finland from the Black Sea in summertime are often accompanied by an open wave which does not occlude.

9.1.3. Inclination of fronts

The geostrophic wind law and hydrostatic equation can be used to approximate the inclination of a front as follows:

$$\tan \alpha = f\,T \frac{V_{g1} - V_{g2}}{g(T_1 - T_2)} \quad ,$$

where $T = (T_1 + T_2) / 2$

In other words, the front is more inclined the greater the difference in temperature, and more upright the greater the difference in wind velocity.

Note: flows are not geostrophic in the proximity of fronts.

If we substitute $f = 10^{-4}$ 1/s, T= 280 K and $g = 10$ m/s^2,

we get $\tan \alpha = 2{,}8 \ 10^{-3} \frac{\nabla V}{\nabla T}$

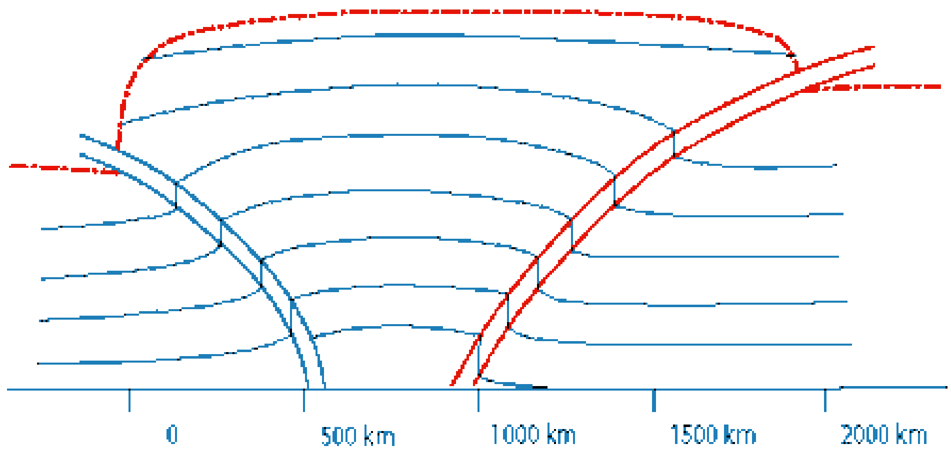
average inclination 1:150

range: 1:50 (steep) and 1:300 (gentle)

the surface front is located in

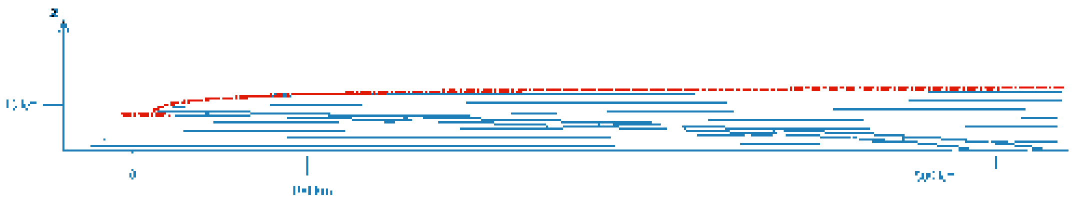
- on the warm side of a 200–1500 km 500 hPa front, or
- at the warm edge of a 500–1000 hPa pressure gradient, and
- on the warm side of a 50–500 km 850 hPa front.

Note: the vertical cross sections of fronts are usually depicted like this:



,

but even in this image the vertical axis has been stretched to approximately three times its real size:



9.1.4. The location of surface fronts

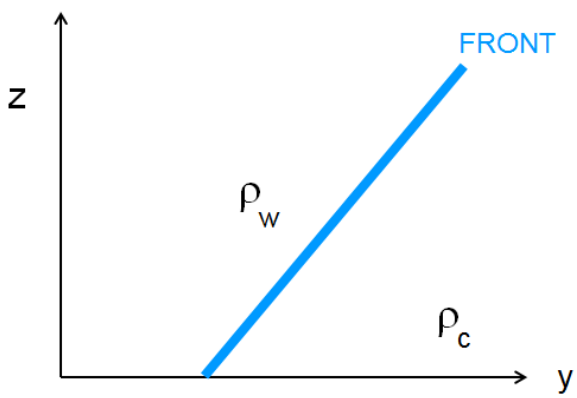
When defining the location of a surface front, it is first approximated from satellite images and numerical fields, mainly temperature and/or equivalent potential temperature at 850 hPa, where the surface does not affect the temperatures. Over mountainous terrain, such as in the Scandes or the Alps, the 700 hPa temperature is used. After that the surface front is defined by using actual surface observations.

The numerical parameters:

The most crucial are different temperature parameters, which are to be looked at 850-700 hPa.

There is also a vorticity maximum in the front, which can be seen as follows (Martin, p.189-190):

Select coordinates: x along the front, y along the the course of the front (perpendicular to x), and z upwards, perpendicular to them both. For the sake of simplicity, let us consider a zero-order front, where the temperature (and density) is discontinuous:



With such a front, the following applies:

$$\frac{dz}{dy} = \frac{(\partial p/\partial y)_c - (\partial p/\partial y)_w}{g(\rho_c - \rho_w)}$$

and $dz/dy > 0$.

This means the pressure gradient is greater on the cold side. The equation above can be written using a geostrophic wind component $u_{(g)}$ aligned with the front:

$$\frac{dz}{dy} = \frac{f\,(\rho_w u_{gw} - \rho_c u_{gc})}{g(\rho_c - \rho_w)}$$

Because $dz/dy > 0$, $u_{(gw)} > u_{(gc)}$, meaning $\partial u_{(g)}/\partial y < 0$, there is positive vorticity present.

There is a local vorticity maximum in the front.

Thermal winds are strong in the upper troposphere in the frontal zone. The component $u_{(g)}$ is roughly equal to the $V_{(g)}$.

At the altitude of jet streams, the front lies on the cold side of the flow in a location where the isotachs are densest and vorticity is high. Because there is a local vorticity maximum in the front, the geostrophic wind law implies that a trough, a pressure gradient change or a col must also be present.

The flow may be anticyclonically curved, in which case the curvature term of the vorticity may be higher than the shear term, and total vorticity might be negative even though it has a local maximum.

On a surface chart the fronts are seen as zones with turning or weakening/strenghtening winds.

Note:

- Not all zones with turning or weakening winds are fronts.
- The estimate is based on geostrophic wind, and surface winds are ageostrophic because of friction.

The observations:

The definition of a front is a zone where temperature changes, so the front is primarily located where the temperature is different on different sides of the front.

However, near the surface its properties as well as lower clouds and changes in the wind may obscure changes in the temperature, or even reverse the expected effects (the so-called masked fronts). The parameters used in these cases are wind and dew point, wind being the more important one.

Additional remarks regarding the location of the surface front:

Polar fronts and air masses are distinct in the free atmosphere, but problematic in the boundary layer and especially its lowest section.

Fronts are drawn in the correct spot on the map because it helps with understanding the weather, so they have to be based on observations, not numerical models.

There are certain rules of thumb regarding a front's location, but many of these criteria often fail to be met due to the effects of the surface.

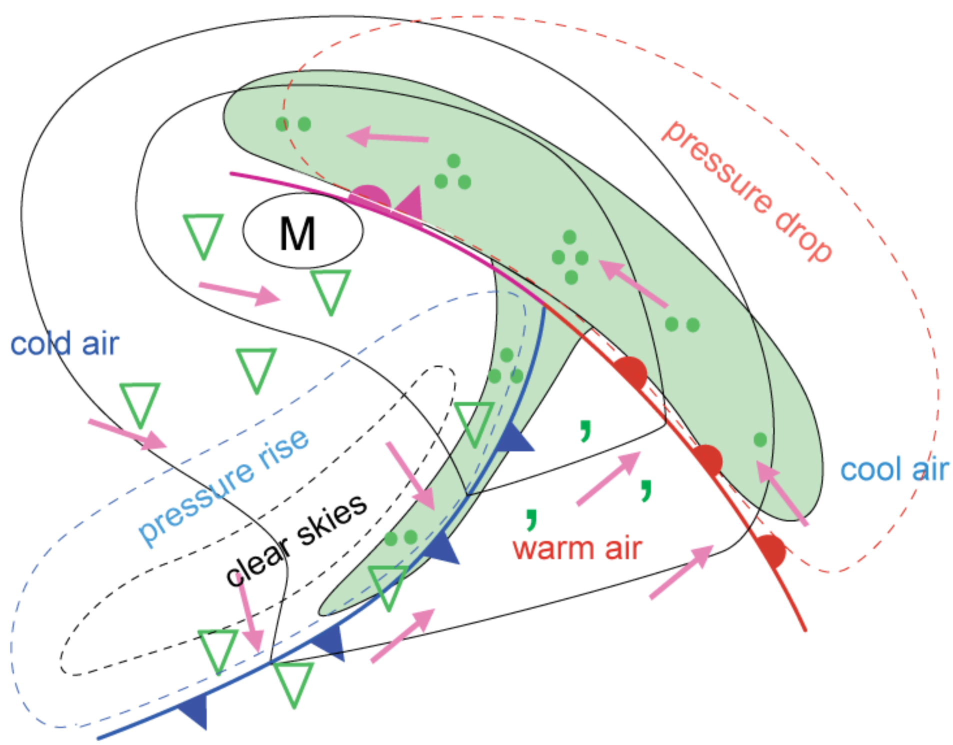
The clearest fronts are found near low pressure centers close to the mature stage of the cyclone.

Occluded surface fronts are often more indistinct that warm or cold fronts.

Some rules regarding finding the location of surface fronts:

- pressure drops ahead of warm and occluded fronts; the largest drops in pressure are often found near the occlusion point
- pressure rises at least awhile behind a cold front (if a trough is following the cold front, pressure will start to drop again)
- the largest and most uniform precipitation areas are found ahead of warm and occluded fronts, and precipitation is often strongest near the occlusion point
- narrow bands of precipitation, continuous and/or showerly, often accompany cold fronts

behind a cold front the skies are often clear where the pressure rises fastest, and a weak ridge will develop there



The fronts drawn on the map are not important by themselves. Rather, their purpose is to explain the weather. Fronts must not be forcibly drawn to correspond to observations they do not match. On the other hand, synoptic scale fronts can include mesoscale features caused by local conditions. Maps of large regions cannot account for them; trying to interpret the weather based on such a map would be confusing.

9.2. Thermal highs and lows

Thermal (warm) lows

In thermal lows, cyclonic development is caused by diabatic heating of the surface. A thermal low develops when the local sensible heat flux is strong.

Thermal lows in Finland are weak, only a few hPa deep. They can be visualized on surface charts by analyzing the pressure field with 1 hPa intervals.

Large-scale thermal lows develop in summer in the tropical regions of warm continents. The most powerful of them is the South Asia monsoon low.

Thermal lows also develop in autumn and early winter over water that is surrounded by land, such as the Great Lakes, the Caspian Sea and the Baltic Sea and its bays.

In Finland, thermal lows typically occur

1. In the Bothnian Bay until the sea freezes, as the warm and humid surface generates convection and showers. If the basic flow is from sea to land, there will also be frictional convergence on the coast, and showers may develop along it.

2. In fair weather with weak winds in spring and summer in southern and western Finland, when a cool sea surrounds the warm land. Afternoons see the development of sea breezes and associated convection roughly 10-100 km inland from the coast.

Thermal (cold) highs

In thermal highs, as in thermal lows, anticyclonic development at the surface is caused by diabatic cooling. In a wintry cold high the outgoing thermal radiation is stronger and the air colder than in its surroundings. A cold high most often develops in a very cold air mass in a weak high or ahead of a weak upper ridge.

Large cold low pressure systems develop over the continents in the northern hemisphere in winter: Greenland, northern Canada, Siberia and Central Asia. They occur so regularly that they can be seen in mean air pressure charts.

A ridge from a strong west Siberian high may reach Finland in winter. In these situations the weather is often very cold.

Thermal highs are usually more intense than thermal lows: as much as dozens of hPa far inland and around 3-10 hPa even in Finland. Thermal highs can also persist for a long time - from days to weeks.

9.3. Orographic pressure anomalies

Orographic pressure anomalies occur as a straight adiabatic air flow crosses a mountain range (height and width at least 1000 km):

- a high pressure ridge tends to form on the windward slope
- a trough tends to form on the lee side

Concerning Finland, the effect of the Scandinavian Mountains on the pressure anomalies of westerly flows is very small.

9.4. Cut-off low

Cut-off lows develop when the amplitude of an upper wave in the polar air mass intensifies and its trough stretches far south and finally breaks off from the basic flow. The result is a cold upper low within the warm air mass, also called a "cold drop".

As a cold upper low forms, a warm upper high or ridge forms to accompany it.

Upper highs and lows are a part of the thermal transfer that goes on between the tropics and polar regions.

Cut-off lows develop inside upper flows. However, in large, deep cold drops cyclonic vorticity will gradually induce a low also to the surface.

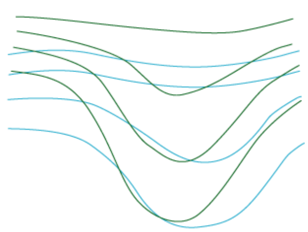
The development of a cut-off low

1. An upper wave amplifies

   The phase difference between the temperature and geopotential waves is such that there is cold advection in the trough and warm advection in the ridge preceding it.

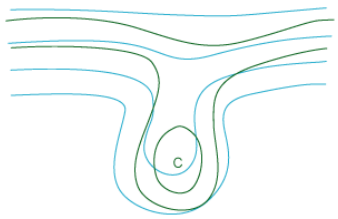
   Tendency equation → the trough deepens

   The turquoise lines in the following schematics represent 500 hPa geopotential and the green lines its temperature:



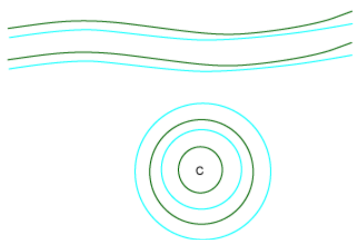
2. A separate low pressure center develops at the bottom of the trough

   As the trough keeps deepening, the warm westerly air on its side approaches the warm air on its eastern side, causing two high pressure ridges to form to the north of the trough.



3. The bottom of the trough becomes detached from the basic flow

   In the final stage of the development the contact between the trough and the basic flow is lost and a closed low pressure, colder than its environment, appears to the south of the basic flow.

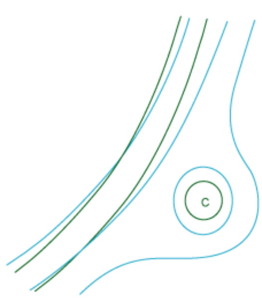


4. The cold drop disappears

   Cold drops may fade out in two ways:

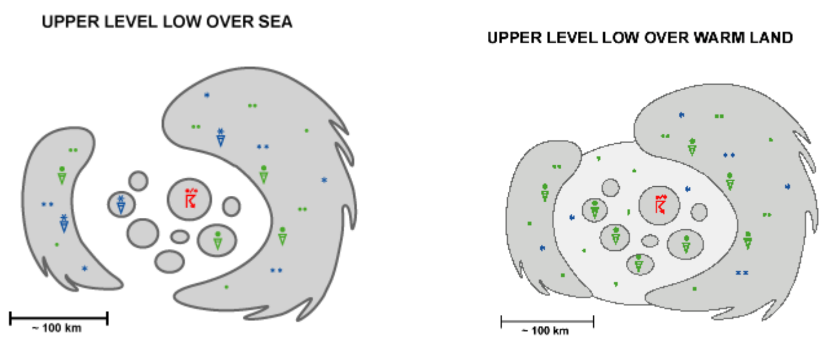
   - usually the drop fades by merging into a intensifying upper wave coming from the west

   - far from regions with pressure gradients a detached cold drop will slowly by the effect of friction



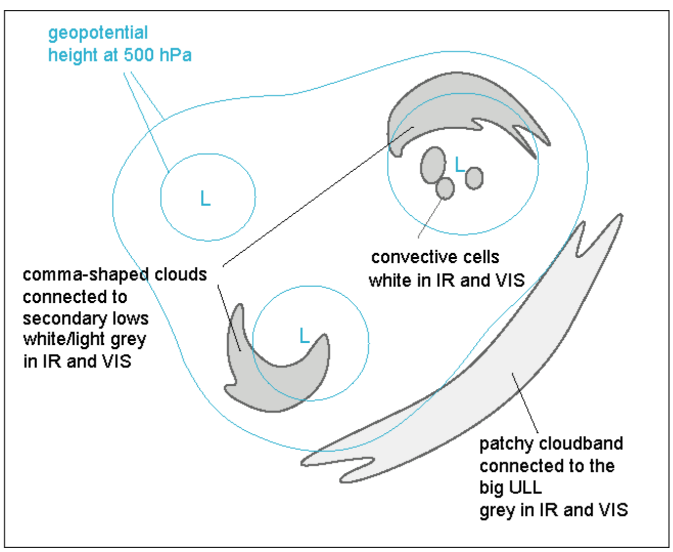
Weather in cut-off lows

Showers and thunder are common in cut-off lows, especially in summer and over sea. In large upper lows there can also occur continuous precipitation in the eastern and western edges of the low.



Baroclinic development

Because there is a temperature gradient at the edges of the cold drop, baroclinic waves may develop there, especially around the eastern edge of a large cut-off low. These waves can undergo baroclinic development, and the result is bunch of secondary lows, which circulate the large low counterclockwise:



9.4. Blocking high

A blocking high is a cut-off high which forms when a warm upper ridge detaches from the basic flow.

Blocking highs are long-lived and widespread, typically several thousand kilometers in width. Cyclones approaching them from the west come to a halt, or they have to go around the blocking on the northern side, which is where they get their name, "blocks".

In a meridional flow, upper lows and upper highs most often detach from the basic flow in pairs; one will be weaker while the other dominates. What typically happens is that a weak upper high or ridge develops ahead of a strong upper low.

The occurrence of blocking highs has two maximums, unrelated to each other: from January through May in the eastern Atlantic, and in January in the Pacific.

Because blocking highs are long-lived and nearly stationary, they often involve longish periods of unchanging weather.

A blocking high contains mid-latitude air masses, descending motion and fairly dry air in the mid- and upper troposphere. The state of the lower troposphere depends on the season.

Typical weather in blocking high –situations in Finland:

Summer

In dry air, there is strong heating near the surface and the temperature rises dry-adiabatically. Diurnal temperature variation can reach 25 °C. Cloudiness is meager; there are either clear skies or low Cu clouds. In humid air, nights will also be warm.

The air is often dry when a blocking high forms, but the strong evaporation will moisten the lower troposphere. Descending motion in the upper troposphere will also weaken gradually, and this results in showers and often finally thunder.

Both day and night temperatures will rise as time goes on, though the night temperatures rise more; the last day of a heat wave resulting from a blocking high is often the hottest.

Autumn

Radiative cooling is strong at night and radiation fogs form easily. As the conditions persist and incoming radiation decreases, the fog will no longer dissipate during the day; instead it rises up and turns into an St or Sc cloud. There is a strong inversion above a shallow moist layer near the surface and diurnal temperature variation is small.

Winter

If the skies are clear, a snow-covered surface will have a strong surface inversion above it, and the temperature is low. St, Sc and Ac clouds as well as radiation fog may develop over snow-free surfaces in above-zero temperature.

Spring

The weather is often dry and clear.

It is typical for all blocking highs that there are flows of warm and humid air in their western and northern borders. On their east and south sides the air is in descending motion and there is fair weather.

Persistent blocking highs exhibit discontinuous retrogression, in which the center of the block moves west. In summer this can have a considerable effect on weather in Finland: the conditions change from dry and sweltering to rainy and chilly.

The disappearance of a blocking high

Short-lived blocking highs (a lifespan of less than a week) blends back in with the mid-latitude air. These blockings are often accompanied with cut-off lows.

A blocking high that lasts about a week will slowly cool, eventually moving off with and merging into the basic flow.

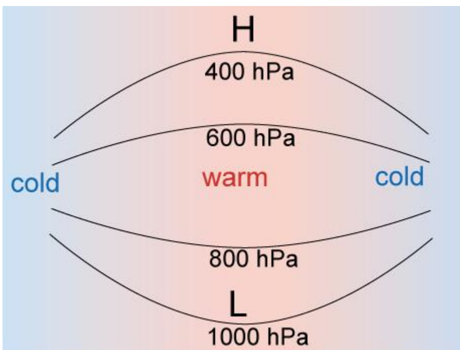
Blocking highs that last over ten days are fueled by moving disturbances that keep it supplied with warm air. When the heat input stops, the high will merge into the basic flow.

9.5. Typical properties of highs and lows

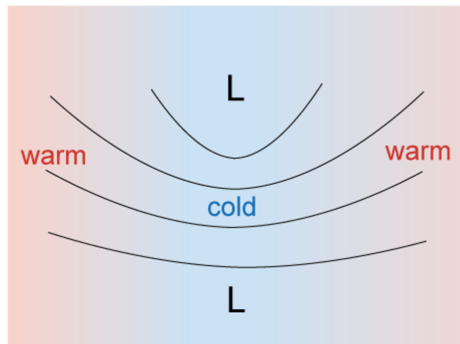
The type/classification of a high or low is not always clear. The nature of different kinds of highs and lows (both stationary and moving) may change, causing one to transform into another. They may also be combinations of different types. However, knowing and recognizing the different types will often help in understanding the weather.

The vertical temperature and pressure distributions of different kinds of highs and lows:

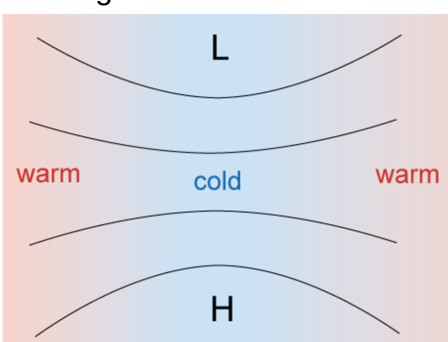
warm low



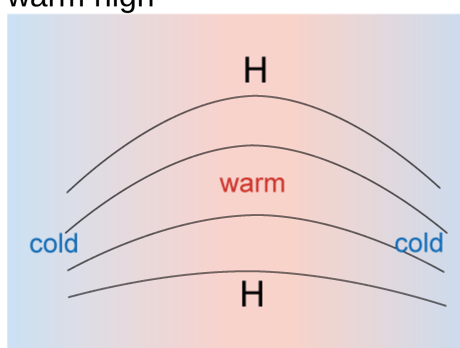
cold low



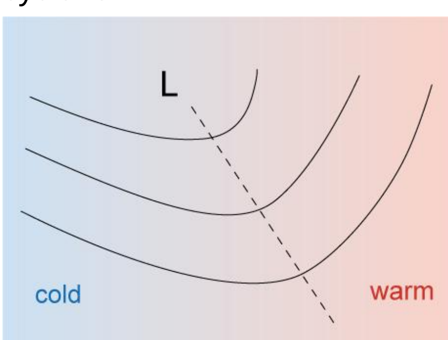
cold high



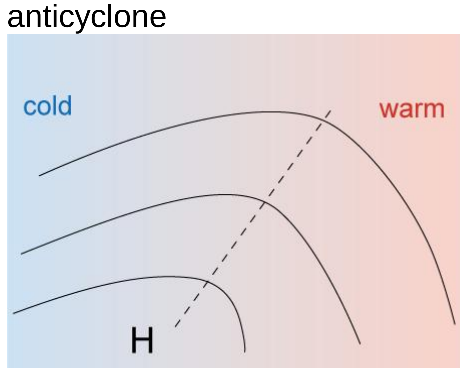
warm high



cyclone



anticyclone



10. Convective systems

10.1. Cold air mass

10.1.1. Trough

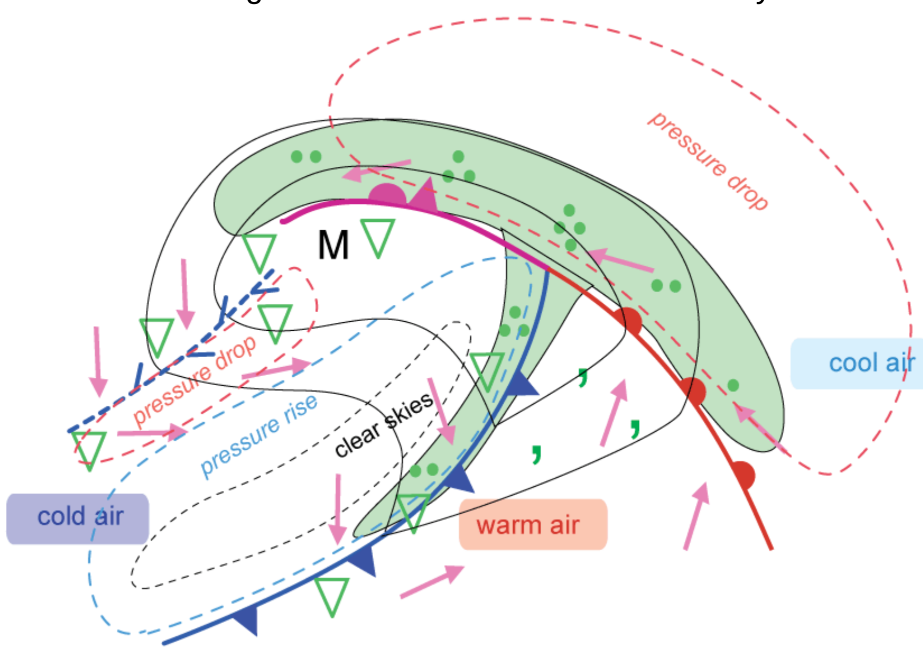
10.1.1.1. Surface trough

Surface troughs are seen as zones of turning winds in surface observations. Because they occur in cold (and therefore slightly stable) air, convection is likely to occur. Depending on humidity, Cu and Cb clouds may form in the trough, but thunder is uncommon.

Surface troughs are often relatively small and travel quickly, so the showers they bring also pass quickly.

Because troughs are located entirely inside cold air masses, they are not accompanied by temperature changes. However, in some cases colder air from the north can flow in from behind a trough, but even then the air cools steadily and there is no barcolinity present.

The most common trough forms behind the cold front of a cyclone:



There might form more than one troughs like this following each other.

The turning of the wind in the end of the trough’s life cycle can be so weak, that it can no longer be clearly seen from surface observations. In summer this kind of trough is likely to be fair in the night, while on the daytime showers will still grow in it.

10.1.1.2. Upper trough

Upper troughs are not visible in the surface observations, but they can be seen in 500 or 300 hPa pressure fields. Sometimes they fall between these main pressure levels, but can be detected from satellite images: they are often accompanied by cyclonically curved comma clouds or a line of Cb clouds. In some cases upper troughs are most easily detected as dark, round spots in water vapor images (so-called water vapor eyes), in which case the clouds are located near the leading edge of the trough.

Showers form in upper troughs. Thunder may also occur in sufficiently intense upper troughs.

10.1.2. Cold air flow

The convection in a cold air flow is shallow convection in the boundary layer: MCC (mesoscale cellular convection). It reaches up to 1-2 km from the surface. Cold air is slightly stable, so convective clouds are easily formed if there is enough humidity, especially over sea. On the top of the boundary layer there is an inversion, which convection can not penetrate.

The clouds are mostly Cu, with small Cb’s embedded. The most common cloud is Cumulus Mediocris. Showers are light, and thunder does not occur.

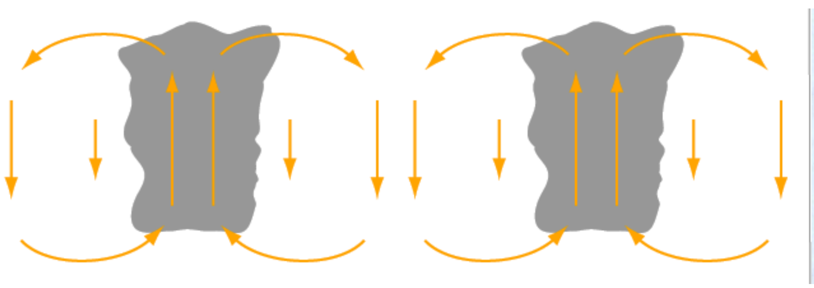
The low convection can form up linearly or hexagonally. Typically, behind a cold front there forms linear convection, which turns gradually into hexagonal one. Linear convection can form cloud streets, which may turn into roll vortices in a strong flow. The rolls are in a roughly 10-20 º angle in regard to the mean wind of the convective layer.

The prerequisite for roll vortices is at least moderate surface wind. They form through two kinds of instability: thermal and dynamical. The thermal instability forms when the cold air flows over a warm surface, while the dynamical one is a result of the combined effect of cold advection and wind shear in the Ekman layer.

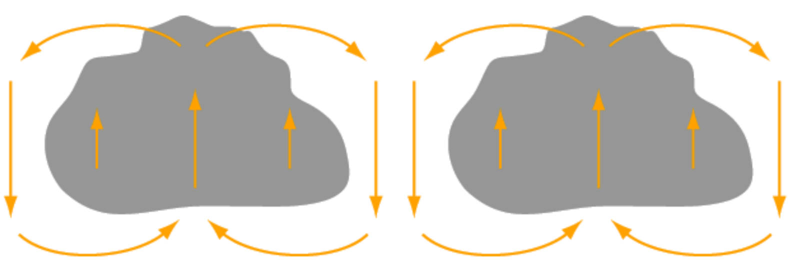
Cloud streets seen from surface and from a satellite:

Hexagonally organized convective clouds form easily over sea. They can be divided into two types:

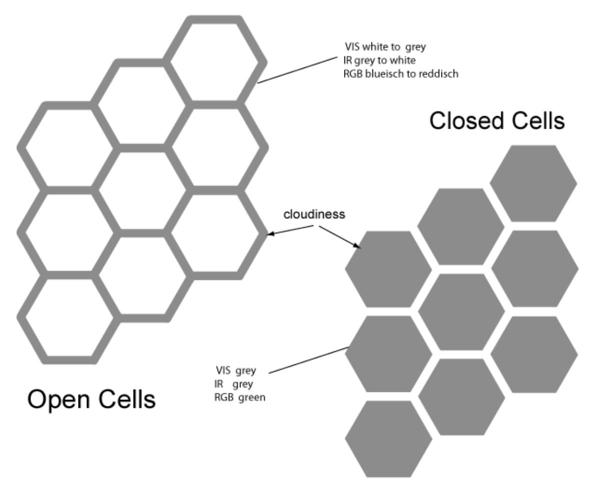
1) occ (open cell convection): large scale vertical motions are descending. In the middle of a cell the air is cloud-free, and it is surrounded by small scale ascending motions.



2) ccc (closed cell convection): large scale vertical motions are ascending. In the middle of a cell there is a convective cloud, and it is surrounded by small scale descending motions.



Seen from below or from above occ looks like separate Cu’s (cloud cover round 4/8), ccc like Sc clouds (cloud cover round 6/8).



10.1.3. Polar low

Polar lows are relatively small cyclonic vortices that may develop over northern seas in wintertime. Polar lows have a warm core on the surface and very cold air higher up, in which case the situation is potentially unstable. In addition, the upper parts of the low require positive vorticity, which means there has to be an upper trough or an upper low there.

A short stretch of baroclinic zone accompanies the warm core in the lower troposphere. It is usually remnant of an old occlusion, sometimes of an arctic front.

The distribution of temperature and positive vorticity advection higher up cause convection, which is intensified by the release of latent heat. The polar low will persist for as long as it gets moisture and heat from the sea, and it will fade soon after arriving over a continent. Polar lows bring dangerously strong, gusty winds and showers - mostly snow showers.

There are two rules of thumb regarding numerical parameters and polar lows: the 10 m wind speed is least 14 m/s, and the 500 hPa temperature -40 ºC or less. Satellite images and numerical fields are often necessary to supplement the scarce surface observations available at sea.

10.2. Warm air mass

10.2.1. Convection in the warm sector over land

If there is sufficient moisture in the warm sector of a cyclone, showers and thunder can develop there in summer. The convective clouds develop in the afteroon following the disappearance of the nightly inversion, which is when the convection begins. In Finland, these kind of showers start occuring around 13:00-15:00, and they reach their peak around 17:00-19:00.

The first signs of convection can be detected in the morning; there will be Altocumulus Castellanus clouds hinting at moisture and ascending motion in the mid-troposphere.

10.2.2. High pressures in the summer over land

In summer, in the afternoon showers and even thunder may develop within high pressures and ridges, similarly as in the warm sector. For this to happen, the air has to be humid enough and without strong descending motion throughout the whole troposphere. Convection in the high pressure system will still not be as strong as in the warm sector.

10.2.3. Spanish Plume

Cyclones have warm conveyor belts that feed warm and humid air into them. On the other hand, very cold air from the stratosphere descends in the wake of cold fronts, and if these two flows meet, the result is potentially unstable air and powerful showers and thunder. The so-called Spanish Plume is a typical example of shower lines that develop ahead of cold fronts.

Spanish Plume: insolation causes strong heating in the Spanish plateau in summertime. When nightly inversion disappears and a cold front moves in from the Atlantic at the same time, explosive convection takes place. Large Cb clusters and MCS’s (mesoscale convective systems) develop in western Europe. These strong thunder clouds can persist long into the night, travelling northeastwards to France and even to British Isles, or to Germany.

11. Numerical parameters in vertical cross sections

11.1. Isentropes

Vertical cross sections of isentropes show stable regions of the troposphere (frontal zones in particular) and potentially unstable or convective regions.

The word "isentrope" refers to entropy: if potential temperature is constant, so is entropy (isentropic = having constant entropy). All reversible adiabatic processes are isentropic.

Entropy: $C_p \cdot \ln\theta$

Potential temperature θ: $\theta = T \cdot \left(\frac{1000}{P}\right)^K$, where $K = \frac{R}{C_p}$

R = gas constant

C_p = heat capacity

In practice the potential temperature θ is the temperature an air parcel would have if it were moved to 1000 hPa adiabatically, meaning without gain or loss of heat.

In thermodynamic diagrams, an adiabat is a curve along which θ is constant; the pressure and temperature of an air parcel may change, but potential temperature does not.

Entropy is conserved in adiabatic processes.

11.2. Adiabatic processes

Adiabatic processes can be diveded into three types:

- dry adiabatic (describes adiabatic changes well if the air parcel is not saturated with water vapor)
- moist adiabatic (air parcel is saturated with water vapor)
- pseudo-adiabatic (all condensed water vapor leaves the parcel)

Since airborne liquid water only contains around 3 % as much amount of heat as water vapor, the pseudo-adiabatic and moist-adiabatic processes are (for practical purposes) identical when it comes to adiabatic cooling.

Equivalent isentropes

> Adiabatic equivalent temperature $T_{(e)}$ is the temperature an air parcel would have if 1) it were to ascend adiabatically so high that all its water vapor condenses, 2) all the water were removed, and 3) the parcel were restored to its original pressure.
>
> Equivalent potential temperature $\theta_{(e)}$ is the potential temperature corresponding to $T_{(e)}$.

When an air parcel is raised adiabatically, the water vapor in it begins to condense at a certain temperature. The altitude (pressure) corresponding to this temperature is called the LCL (lifted condensation level). Above it the process becomes moist-adiabatic or pseudo-moist-adiabatic.

11.3. Tropospheric stability and potential temperature

When an air parcel is lifted under labile conditions, it will continue its ascent. Under neutral conditions the parcel will remain stationary, and under stable conditions it will head back down.

On the other hand, when an air parcel rises dry-adiabatically, its potential temperature is conserved. If, however, the potential temperature of the parcel's environment decreases at the same time, the air will cool faster around the parcel than it does elsewhere. In that case the situation is labile.

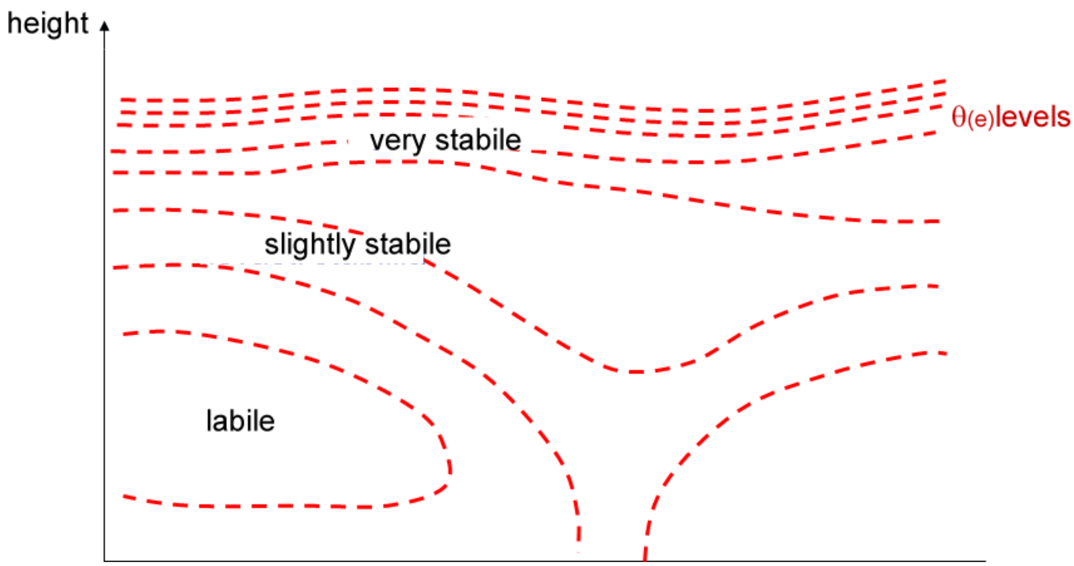
In other words: the conditions are

- labile, when θ decreases with height
- neutral, when θ is constant
- stabile, when θ increases with height

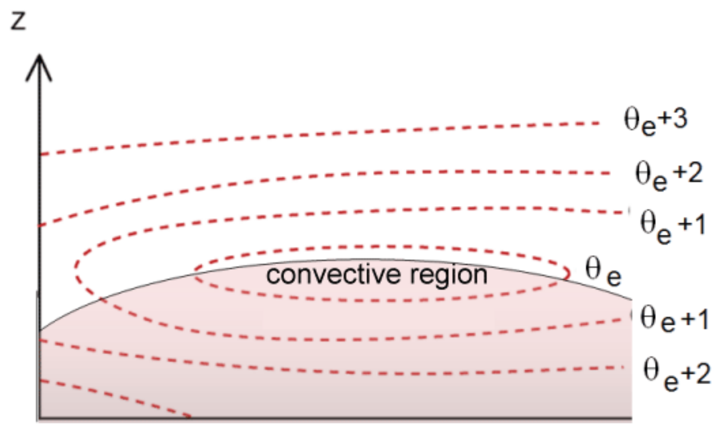
According to the definition, $\theta = T \cdot (1000/P)^{\kappa}$, meaning potential temperature is inversely proportional to pressure. This means isentropic surfaces slope in an opposite direction from pressure surfaces: down toward warm air and up toward cold air. In other words, pressure and temperature increase when descending along an adiabat, and decrease while ascending along it.

11.4. Isentropes in vertical cross sections

The faster θ(e) increases with height, the more stable the situation.The stratosphere is always much more stable than the troposphere, so the tropopause appears in isentropic cross sections as a boundary of the stable region. Generally speaking, the distance between isentropes is a measure of stability.



θ(e) decreases with height under convective or conditionally unstable conditions:

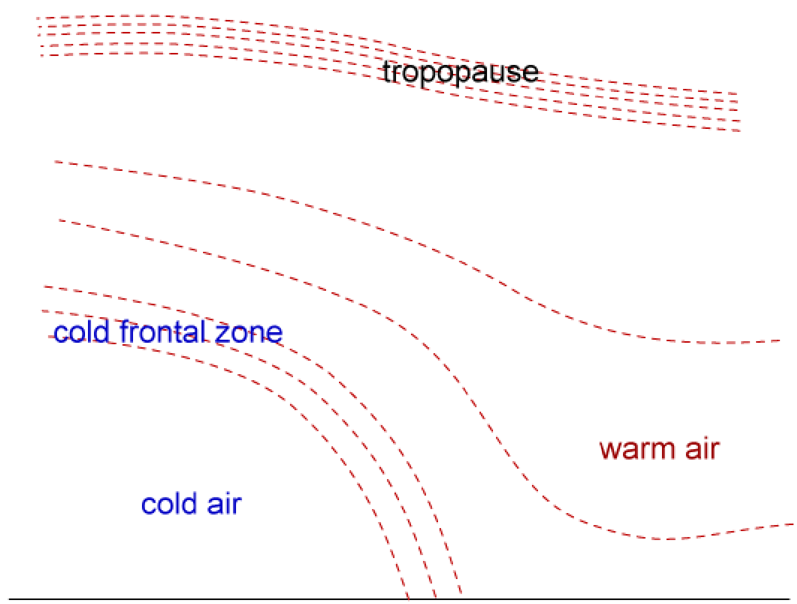


11.5. Isentropes and baroclinic zones

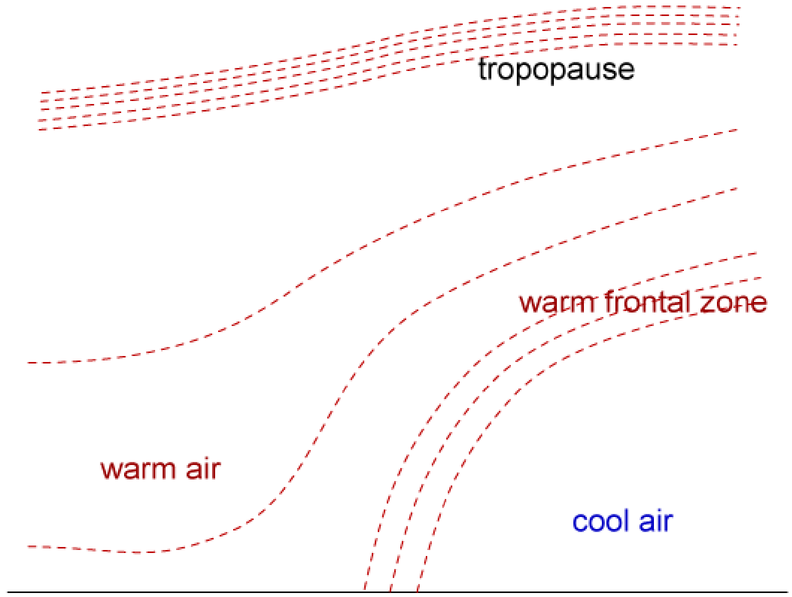
Baroclinic zones are the most stable parts of the troposphere. Vertical cross sections are convenient in analyzing them, because

- isentropic surfaces are nearly parallel to frontal zones
- stability is greater in fronts than in their environment, meaning isentropes are closer to each other there than elsewhere in the troposphere (they are sometimes called crowding zones because of this)
- one can see fronts that extend throughout the troposphere, upper level fronts and low-lying arctic fronts

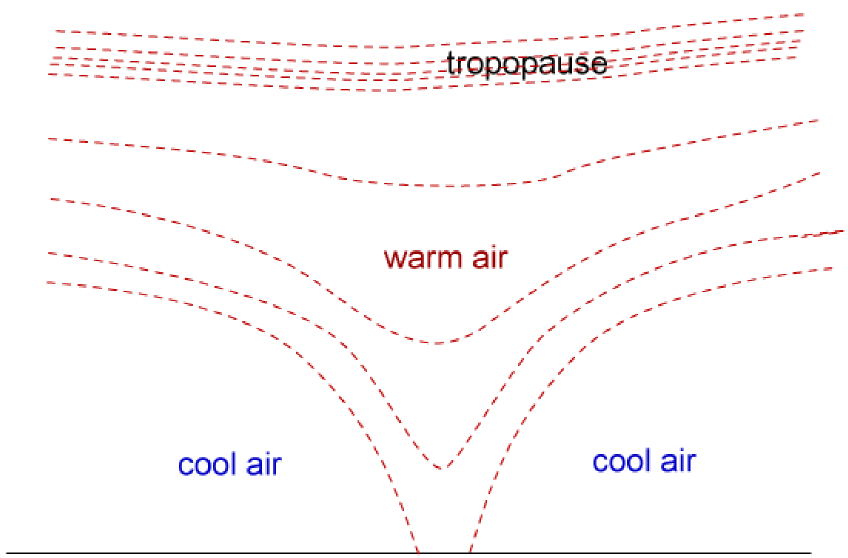
Cold front in isentropic cross sections:

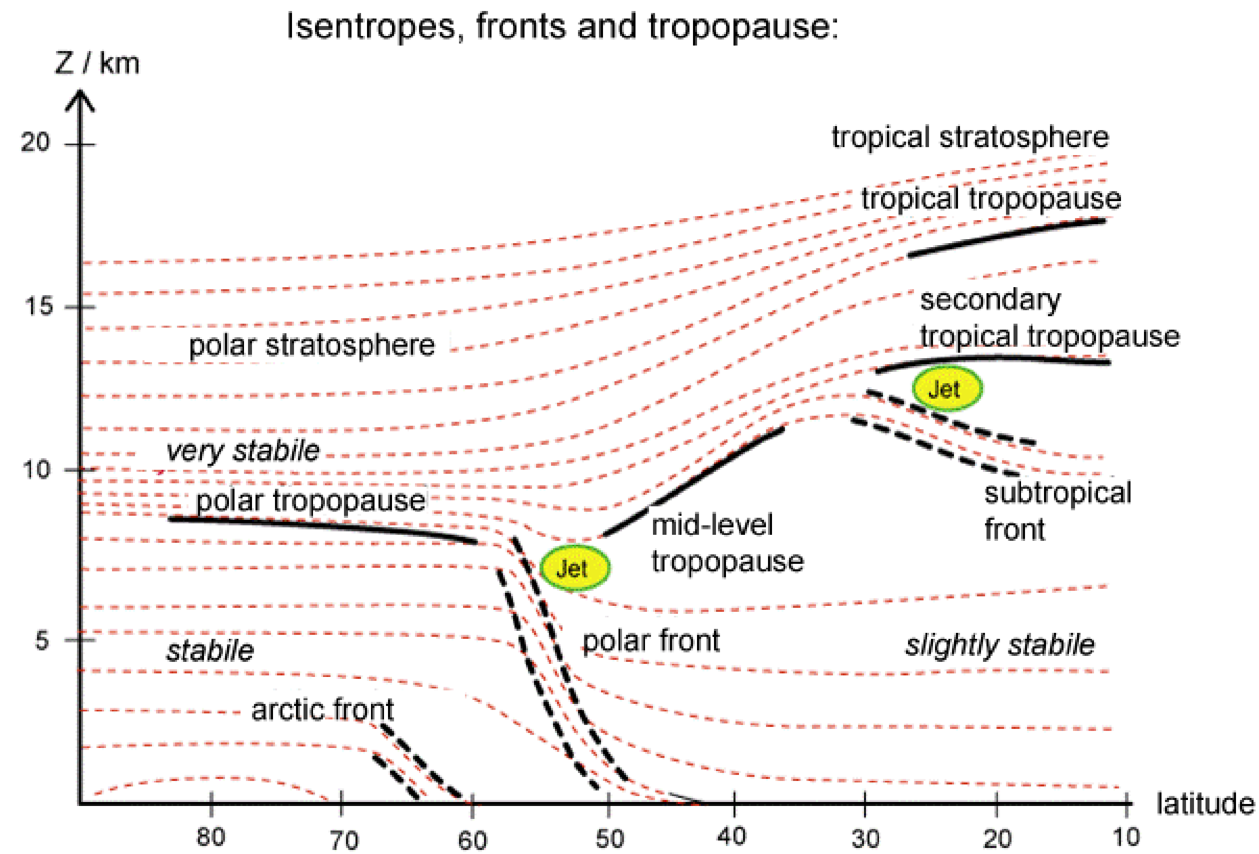


Warm front in isentropic cross sections:



Occlusion in isentropic cross sections:



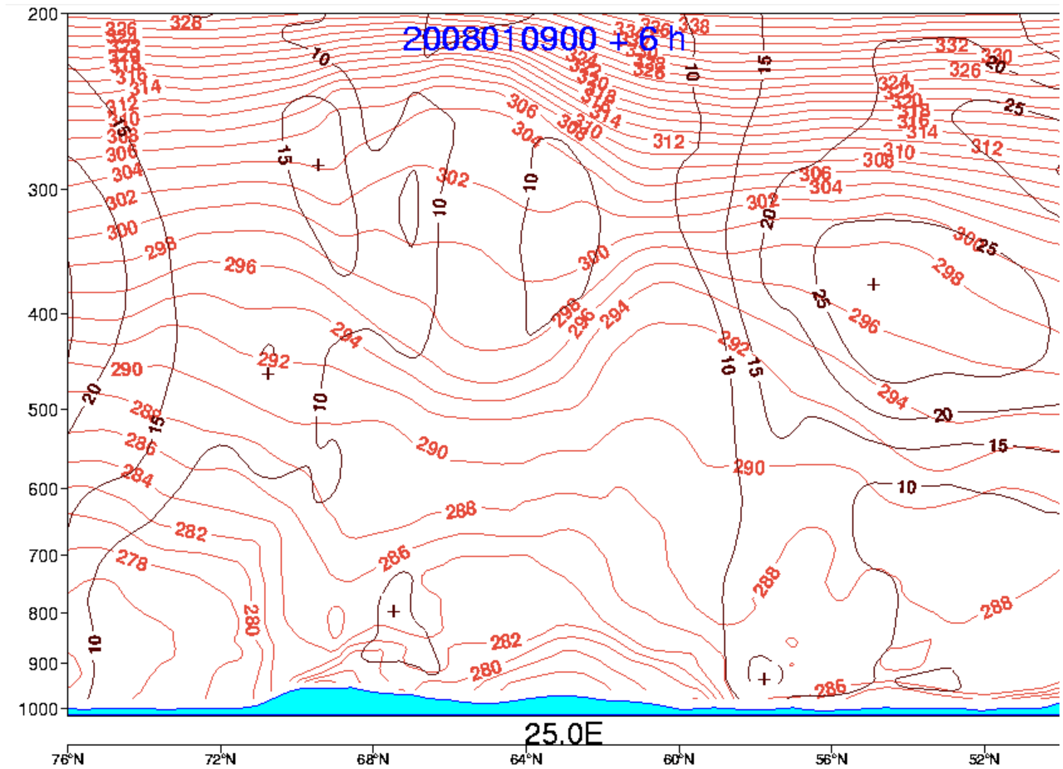


When analyzing isentropes, one has to keep in mind that the atmosphere is not fully adiabatic, especially in the boundary layer or convective situations. Isentropes are vague in over-adiabatic conditions and discontinuous in the proximity of strong diabatic processes.

Usually some additional parameter accompanying isentropes is given in vertical cross sections of numerical parameters:

- isotachs
- omega
- divergence
- relative humidity
- vorticity advection
- temperature advection
- temperature
- potential vorticity

An example with isotaches:



The picture above clearly shows how an upper tropospheric wind maximum may be located at a different height than 300 hPa, which is where one typically looks for jet streams. 400 hPa would often be better in the case of cold air masses, while 100 or 200 hPa would be better for warm air masses.

12. Mid-latitude cyclones

12.1. The quasi-geostrophic cyclone model

12.1.1. Baroclinic instability

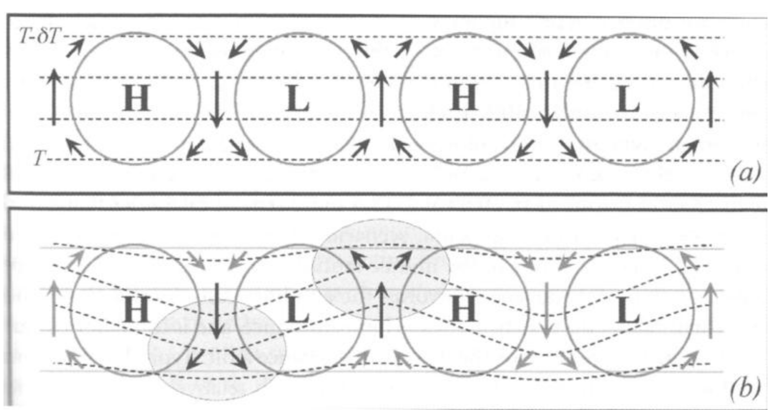
Assume the flow in the mid-latitudes to be zonal and that it follows the thermal wind law. This means that isobars and isotherms are parallel somewhere in the upper troposphere. When a wave disturbance of equal velocity with the basic flow is introduced into it, meridional movements are generated in the flow. These movements bring warm advection downstream of the trough's axis and cold advection upstream of the trough's axis (Martin, figure 8.5.).

This generates a temperature wave that trails the geopotential wave by one quarter of the wavelength.The wave will intensify if both temperature anomalies intensify, and the kinetic energy of the waves increases

If a system with temperature gradient ends up in a state of less potential energy (Martin, figure 8.3.), some of the potential energy turns into kinetic energy (APE = available potential energy).

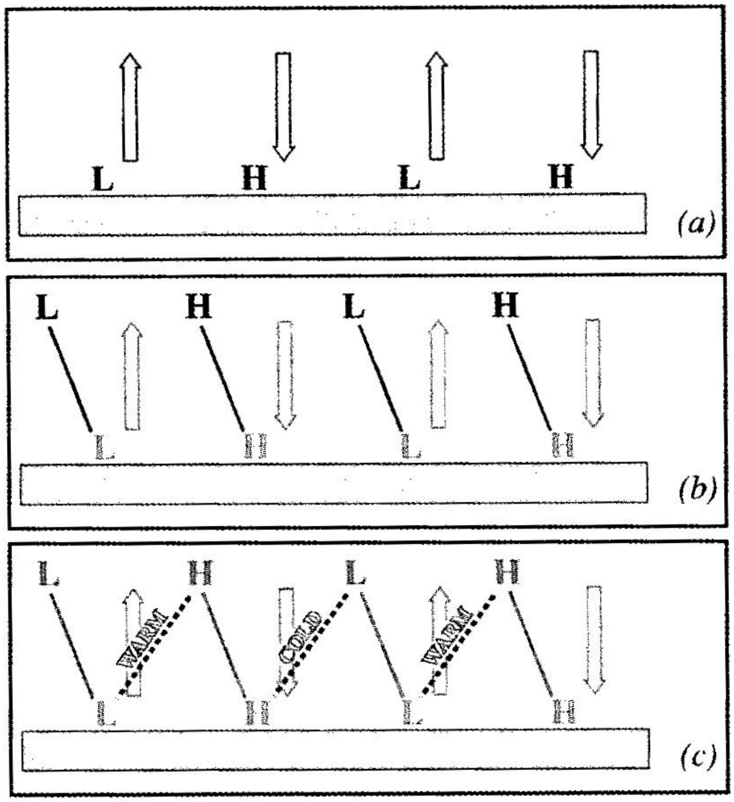
When the APE of the basic flow turns into kinetic energy in the wave disturbance, the flow is unstable with respect to the disturbance.

Low and high pressures in the mid-latitudes are wavelike phenomena. The isobars on surface charts form a series of highs and lows:



H and L: surface highs and lows, arrows: geostrophic wind dashed lines: isotherms, shaded areas: regions of frontal zone development

For the air pressure to remain lower inside a low pressure than its environment, air must leave form the air column above it, and correspondingly the air column above a high pressure must draw in more air. A series of highs and lows is therefore a line of areas where the pressure rises and falls:



(Martin, figure 8.7.)

In the mid-troposphere ascending motion lies downstream of the trough and descending motion upstream (see Martin, figure 8.3.).

Thickness is proportionate to the mean temperature of the air column, so in the upper troposphere the geopotential minimum lies over cold air, and the maximum over warm air. The surface low, on the other hand, is at the peak of the warm sector. This means the axis of a developing geopotential wave tilts backwards wirth height, whereas the axis of a temperature wave tilts forward.

Air rises in warm columns and descends in cold ones. This causes direct circulation to convert the background baroclinity's APE into kinetic energy in developing mid-latitude disturbances.

The zonal baroclinic instability of the basic flow is greatest for waves with a wavelength of 3000-4500 km.

Meridional shear produces thermal troughs and ridges. Diffluence to the northeast or southwest of highs and lows strengthens the temperature gradient. This causes frontogenesis in these areas.
→
Frontogenesis is a consequence of cyclogenesis (not cause like in the Norwegian model!).

In cyclogenesis, baroclinic instability requires there to be considerable shear in the basic flow. The temperature gradient produced by the thermal wind, together with the wave in the upper flow, create an environment where vorticity advection initiates ascending motion, which then generates and intensifies a surface low.

The ascending motion begins at the same time as a thermal ridge develops downstream of the deepening surface low. The circulation accompanying the low shapes the baroclinic zone through advection.

12.1.2. Cyclogenesis and the tendency equation

Cyclogenesis refers to the initiation and development of a surface low. The intensity of a low is often measured by the pressure drop taking place in the low's center.

The same process generates vorticity near the surface, so cyclogenesis can also be considered as an increase in lower tropospheric vorticity.

On the other hand, production of vorticity is related to divergence and vertical motions, which gives us the pressure tendency equation for surface pressure changes with time:

$$\frac{\partial p_s}{\partial t} \sim -\int_0^s (\nabla \cdot \vec{V})\, \partial p$$

A change in surface pressure at a certain location therefore depends on the divergence in the air column above it: pressure drops if divergence is present, and rises is there is convergence instead.

Because divergence cannot be measured very precisely, the equation is converted into a more usable form with the quasi-geostrophic vorticity equation and the quasi-geostrophic thermodynamic energy equation. The result is the quasi-geostrophic height tendency equation:

geopotential tendency χ:

$$\chi \equiv \frac{\partial \Phi}{\partial t} \approx g \frac{\partial z}{\partial t}$$

tendency equation:

$$\underbrace{\left(\nabla^2 + \frac{f_o^2}{\sigma}\frac{\partial^2}{\partial p^2}\right)\chi}_{\substack{\text{A} \\ \text{change in height}}} = \underbrace{-f_o \mathbf{V}_g \cdot \nabla \left(\frac{1}{f_o}\nabla^2 \Phi + f\right)}_{\substack{\text{B} \\ \text{vorticity advection}}} - \underbrace{\frac{f_o^2}{\sigma}\frac{\partial}{\partial p}\left(-\mathbf{V}_g \cdot \nabla \frac{\partial \Phi}{\partial p}\right)}_{\substack{\text{C} \\ \text{differential} \\ \text{temperature} \\ \text{advection}}}$$

$\sigma > 0$ in statically stabile atmosphere

$f_o = 2\,\Omega \sin \phi_o$

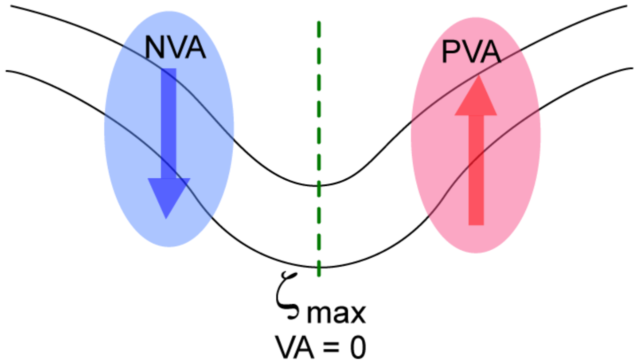
Vorticity advection propagates the wave forward and differential temperature advection either strengthens or weakens the wave.

In developing cyclones, there is warm advection ahead of the surface low. It is strongest in the lower troposphere and weakens as pressure drops (height increases). In this case the differential temperature advection is negative and pressure rises: a ridge develops in the mid-troposphere ahead of the surface low, near the warm front. Correspondingly, a trough in the mid-troposphere develops behind a cold front, since cold advection is stronger in the lower troposphere and weakens with height.

12.1.3. Cyclogenesis and the omega equation

Nearly all cyclogenesis begins as a disturbance (wave) in the upper flow. This disturbance shows as a maximum of relative vorticity.The disturbance is strongest in the mid- and upper troposphere, where the geostrophic wind is also strongest. This being the case, there is positive vorticity advection downstream of the disturbance, and negative vorticity advection upstream of it. Both increase with height.

According to the quasi-geostrophic omega equation, there is ascending motion downstream of a trough and descending motion upstream of it:



12.1.4. Q-vectors

The positive vorticty advection accompanying thermal winds is related to ascending motion, and negative vorticity advection to descending motion. This connection can be studied in compact form with the so-called Q-vectors, which describe the transformation rates of potential temperature gradients with geostrophic winds.

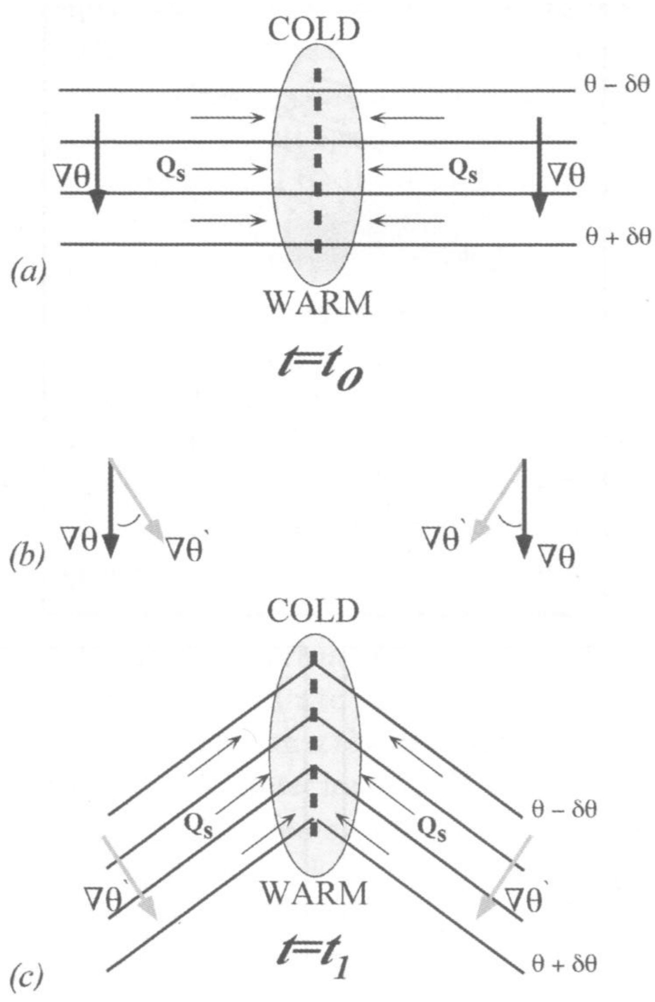
If the Coriolis parameter is assumed constant, the omega equation can be written as follows in the z-coordinate system:

$$\left(N^2 \nabla^2 + f_o^2 \frac{\partial^2}{\partial z^2} \right) w = 2 \nabla \cdot \overline{Q} \,,$$

where N is the static stability parameter.

Ascending motion w can therefore derived from the divergence of Q-vectors.

The Q-vector distribution changes the temperature field, producing a thermal ridge near the surface low and a thermal trough downstream of it:



(Martin, figure 8.21.)

Here Q(s) is a Q-vector component parallel to the isentropes.

a) straight baroclinic zone with convergence of Q(s)
b) ascending motion and curving of the dθ
c) thermal ridge

Adiabatic cooling accompanies ascending motion. The ascent maximum is located at the top of a warm ridge, which the maximum erodes by its presence. However, the release of latent heat compensates for this effect.

12.2. Large-scale flows and the Shapiro-Keyser model

In the Shapiro-Keyser model powerful Atlantic cyclones are studied with the help of numerical simulations.

In 1990, Shapiro and Keyser formulated the so-called T-bone model, in which the warm front of a developing cyclone stretches backward at the same time when the cold front detaches from it. They called this extension of the warm front a "back bent warm front".

Structures like this had been known since the 1940s, but not much attention had been paid to them until numerical simulations made it possible to study them more closely.

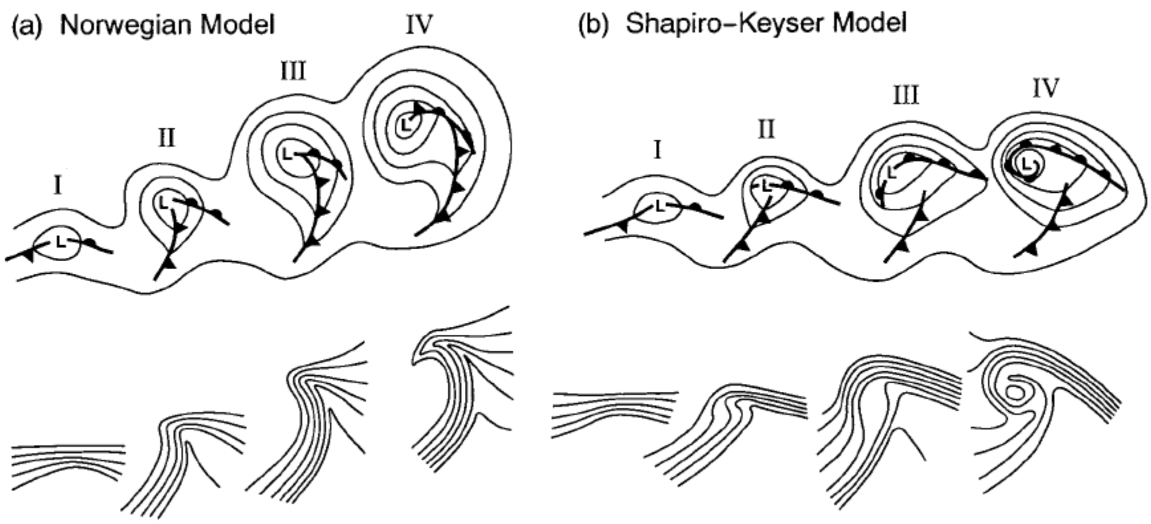
Bent-back fronts have also been called by the following names:

- bent-back occlusion
- extension of an occlusion backwards
- bent-back warm front
- bent-back front

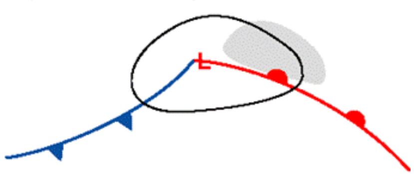
According to the Norwegian model, what happens in the occlusion process is that the warm sector becomes narrower and starts to disappear from the surface up. The center of the low starts to fill and its movement slows down. The cyclonic flow is still strong, however, and as warm air ascends, a cloud spiral develops. Fronts are still clear in the early stages of the occlusion, but especially the warm front becomes increasingly undefined as the occlusion proceeds.

The development of a low in a Shapiro-Keyser cyclone begins as a wave in the polar front. As time passes, the temperature gradient weakens near the center of the low, in the vicinity of the cold front. At the same time, the temperature gradient accompanying the warm front intensifies, and a temperature gradient also forms behind the low, causing the warm front to stretch behind the low. It is nearly perpendicular to the cold front, hence the term T-bone. After this the rear part of the warm front coils around the low and forms a warm seclusion there.

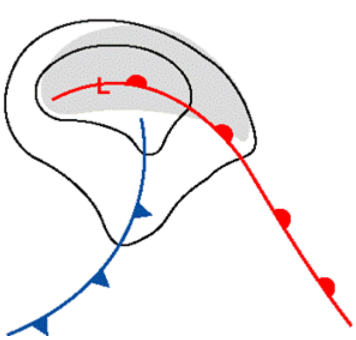
The heaviest precipitation happens on the cold side of the bent-back front, and heavy showers occur further in the cold front. Strong, gusty winds near the end of the bent-back front are typical for T-bone lows at sea ("poisonous tail of a cyclone"). Relatively large amounts of precipitation are typical of continental T-bones.



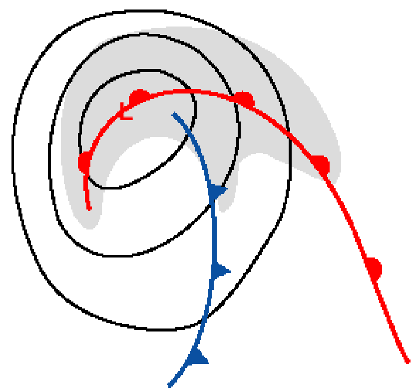
Shapiro-Keyser I: the forming of low pressure



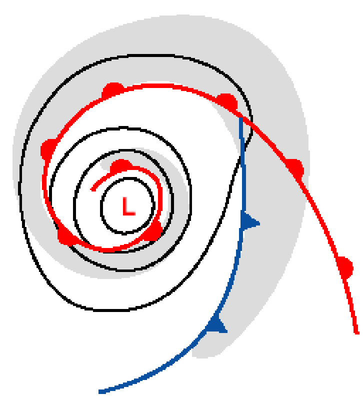
Shapiro-Keyser II: frontal fracture

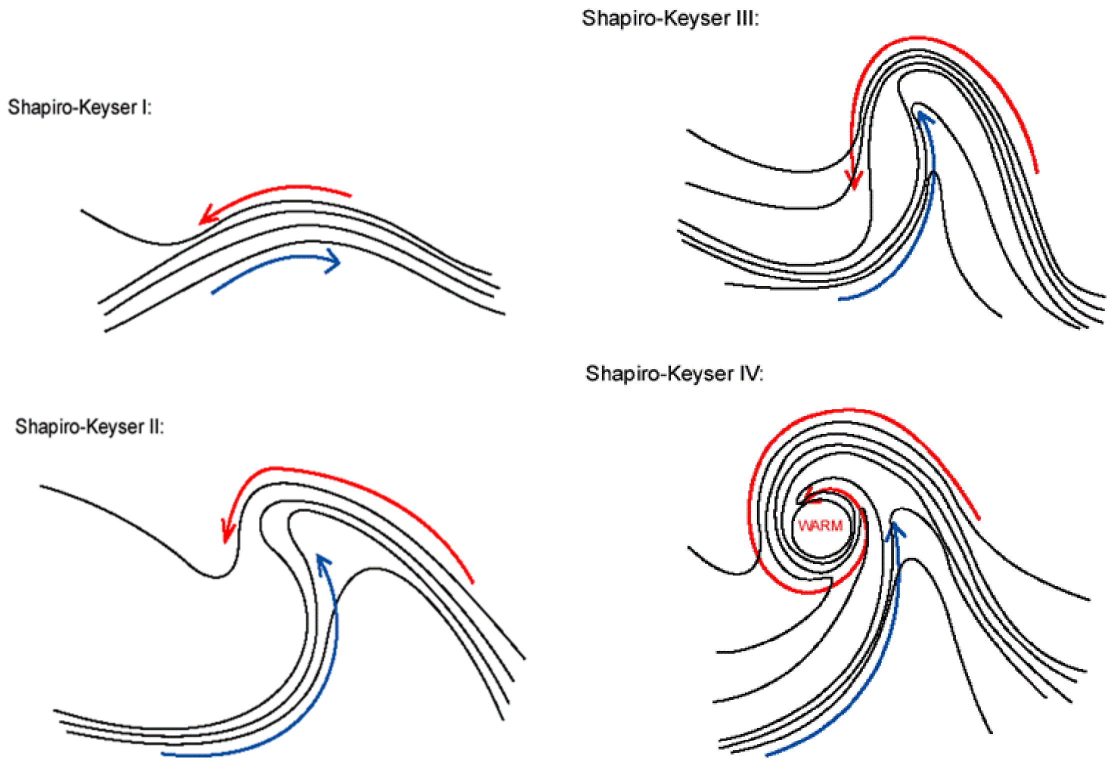


Shapiro-Keyser III: bent back warm front and frontal T-bone



Shapiro-Keyser IV: warm-core frontal seclusion





The influence of the basic flow

When a low pressure system arrives in a diffluent ridge with a large amplitude, the low stretches meridionally. A strong meridional cold front and a weak warm front develop inside it, and it occludes consistently with the Norwegian model.

This development is typical in westerly flow for lows in the eastern parts of the North Atlantic Ocean, which is often the exit region of jet streams.

Diffluence effects in two ways:

- Vorticity anomalies and fronts stretch meridionally. The axes of stretching are parallel with the isentropes, and frontogenesis occurs on both sides of the warm sector.

- When a low starts to move more to the north and less to the east, the cold
  air traveling east begins to encircle the center of the low, forming an occlusion.

When a cyclone arrivews at confluent zonal flow with weak amplitude, it stretches zonally, causing a strong, zonal warm front and a weak cold front to develop inside it. Such a low is of the T-bone variety, and its development follows the Shapiro-Keyser model; it develops toward a warm seclusion via the bent-back warm front and the detached cold front.

This development is typical in the western parts of the North Atlantic Ocean, which is often the entrance region of jet streams. The T-bone structure is a sign that the isentropes are rapidly changing direction from the meridional of the cold front to the zonal of the warm front. The isentropes and axes of stretching are perpendicular to each other in the zone between fronts. This causes deformation to drive the isentropes further apart and frontolysis to occur in the area. On the other hand a bent-back warm front prevents the cold air from coiling around the center of the low, which is moving eastward.

T-bone development can change and start following the Norwegian model if a low travels over the Atlantic from a confluent region into a diffluent one.

12.3. Latent heat and rapid cyclogenesis

Rapid Cyclogenesis = Explosive Cyclogenesis

Properties of rapid cyclogenesis:

- a low deepens at least 24 hPa in 24 hours
- often develops in the warm western parts of the Gulf Stream
- causes considerable damage: storm winds, torrential rains and floods, thunder

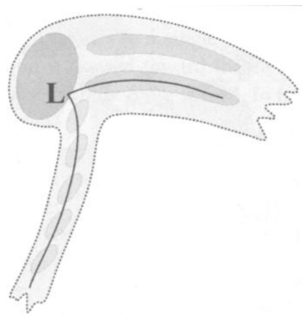
Rapid cyclogenesis is related to particularly strong ascending motions. In the omega equation the stability parameter is the multiplier of ω:

$$\sigma\left(\nabla^2 + \frac{f_o^2}{\sigma}\frac{\partial^2}{\partial p^2}\right)\omega = f_o \frac{\partial}{\partial p}\left[\bar{V}_g \cdot \nabla(\zeta_g + f)\right] + \nabla^2\left[\bar{V}_g \cdot \nabla\left(-\frac{\partial \phi}{\partial p}\right)\right]$$

When the other factors remain constant, the strength of the ascending motion depends on the stability parameter σ. The smaller the stability, the stronger the ascending motion.

The parts of a cyclone where precipitation is strongest are typically:

- a broad precipitation area near the center of the low
- precipitation bands accompanying the warm front
- convective precipitation accompanying the cold front



(Martin, figure 8.13.)

The development is fastest when an area of strong precipitation forms north and west of the center of an eastbound cyclone. In those conditions the release of latent heat

- increases the energy of the system
- strengthens vertical motions by decreasing static stability locally
- affects the large-scale structure and dynamics so that the cyclogenesis intensifies ("self-development")

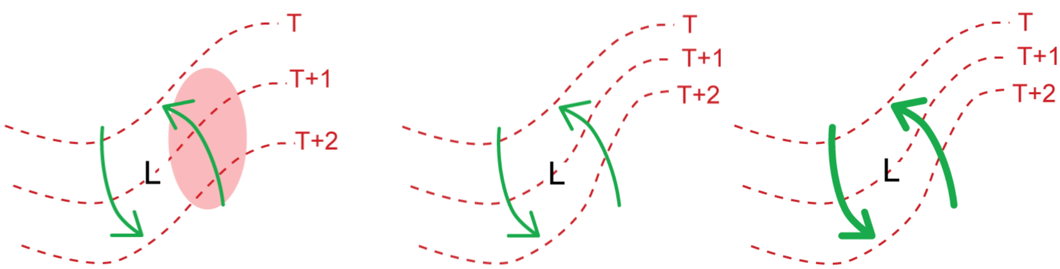
Latent heat is released when water vapor condenses into water droplets. Heat is molecular movement, and there is a great deal less of it in water than in vapor. The phase changes of water involve large amounts of energy.

12.4. Self-development

Consider a cyclone traveling in a westerly flow. The poleward-blowing boundary layer wind on the east side of the cyclone, together with the ascending air, heats the lower troposphere ahead of the warm front.

The heating is partly adiabatic: warm advection, partly diabatic: latent heat release in ascending air. The heating causes the temperature gradient in the lower troposphere to intensify, which strengthens warm advection. The increase in warm advection in turn increases ascending motion.

A stronger ascending motion results in an increase of baroclinic potential energy - a stronger cyclone. The increase in circulation inside the cyclone causes a positive feedback, meaning the disturbance intensifies itself.



a b c

a) The temperature gradient is uniform at first. Heat fluxes heat the lower troposphere in the shaded area.

b) The heating causes the temperature gradient to tighten on the east side of the low.

c) More intense low has stronger winds in the lower troposphere, and they increase warm advection even further.

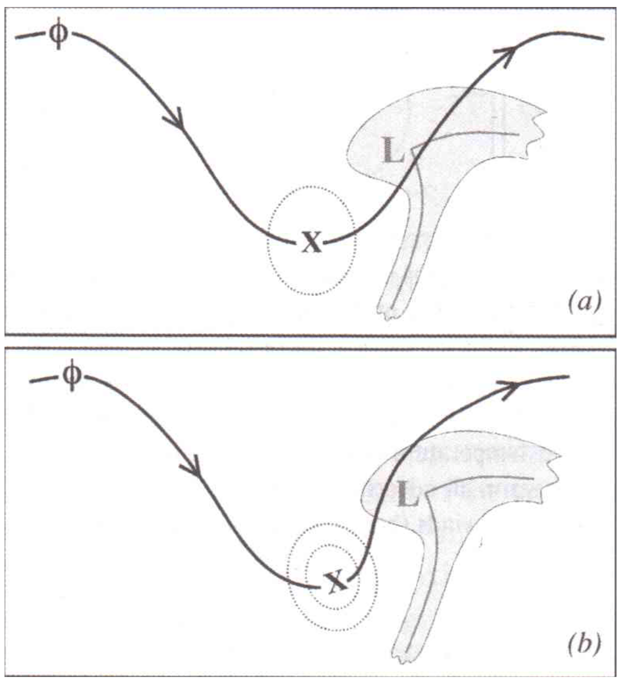
Something to note about figure B: circulation also increases cold advection west of the low, but this effect is toned down by the absence of diabatic factors.

On the larger scale, the release of latent heat results in a positive thickness anomaly immediately downstream of the trough's axis. This causes the geopotential to rise in the mid- and upper troposphere, which generates a high pressure ridge over the area where latent heat is released.

The positive vorticity advection near the front edge of the trough shifts the wave downstream. Because a diabatic ridge is developing there at the same time, the wavelength shortens.

The shortening wavelength causes a significant increase in cyclonic vorticity advection, which further increases ascending motion downstream of the trough's axis.

Increased ascending motion intensifies the low and produces more latent heat ahead of the trough, and this reduces the wavelength of the upper wave. This results in positive feedback.



a) black wave: an upper wave (500 hPa geopotential)
   X = vorticity maximum
   L = center of a surface low

b) The release of latent heat alters the geopotential by forming a small-scale ridge at the front edge of the trough
   →
   greater curvature vorticity near the trough's axis

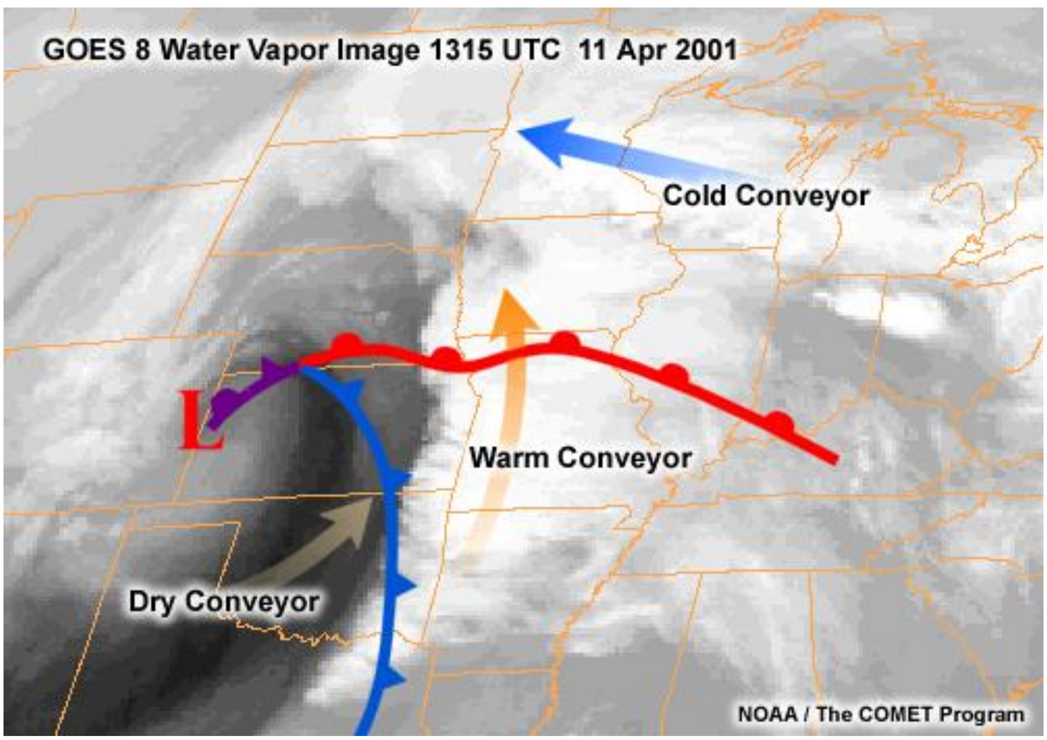
12.5. Conveyor belts

The quasi-geostrophic model uses Euler co-ordinates, in which the cyclone moves through a fixed coordinate system. The model of conveyor belts presented by Carlson (1980) examines cyclones with Lagrangian coordinates, where cyclone travels with the coordinate system. The conveyor belt model studies air flows in relation to a low pressure system's center and fronts.

Conveyor belts are relatively narrow air flows that travel along isentropic surfaces. These so-called relative flows offer a perspective on the clouds and precipitation associated with cyclones that differs from the quasi-geostrophic model.

There are three kinds of relative flows:

- warm conveyor belts
- cold conveyor belts
- dry intrusions



In addition to these three basical flows, conveyor belts have been subdivided further:

Schultz 2001: two cold conveyor belts (CCB)

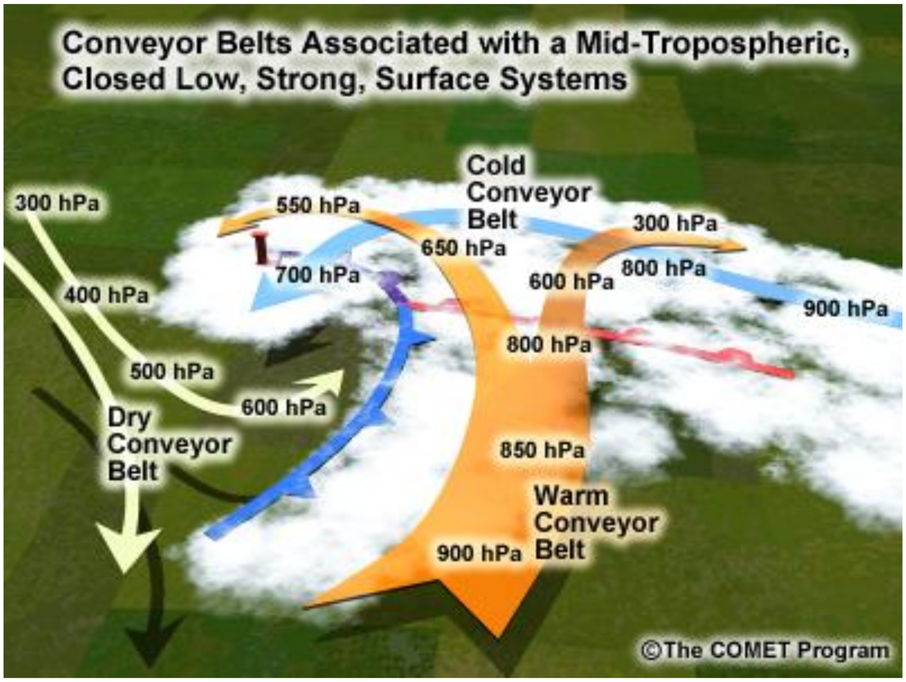
- Anticyclonic CCB originates in the mid-troposphere and rises toward the peak of the warm sector ahead of the front. Once there, it turns clockwise, separating from the low pressure system in the process.

- Cyclonic CCB is a lower-tropospheric flow, which takes a counterclockwise route around the center of the low.

In a strong, deepening lows there is only cyclonic CCB. Cloudiness and precipitation are mostly related to the humid, rising warm conveyor belt.

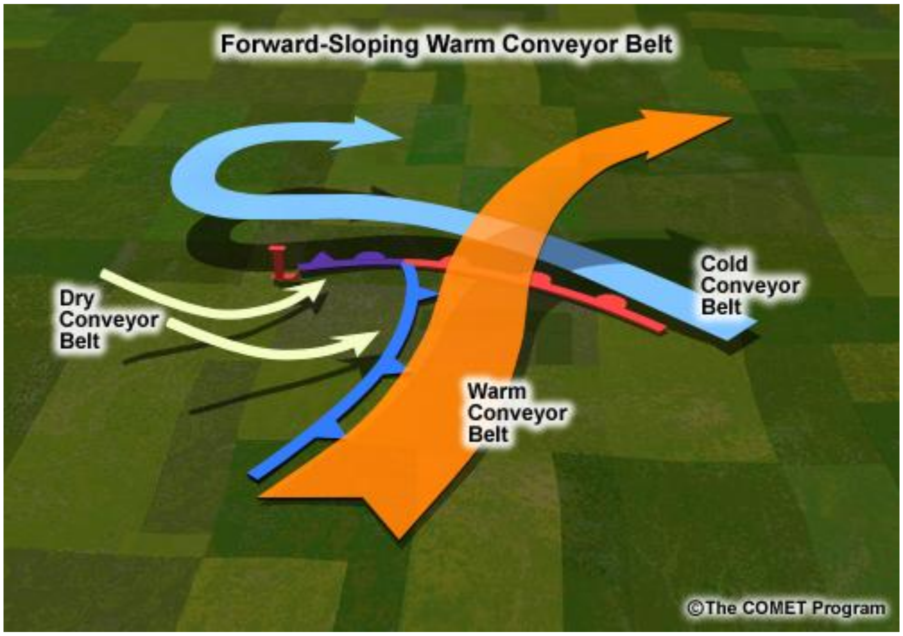
There are clear boundaries between the CCBs in lows with well-defined and strong warm fronts. Conversely, if the warm front is weak, the boundary between the CCBs is wider and less definite.

It has also been proposed (by The COMET Program), that warm conveyor belts split into two branches in occluded cyclones:

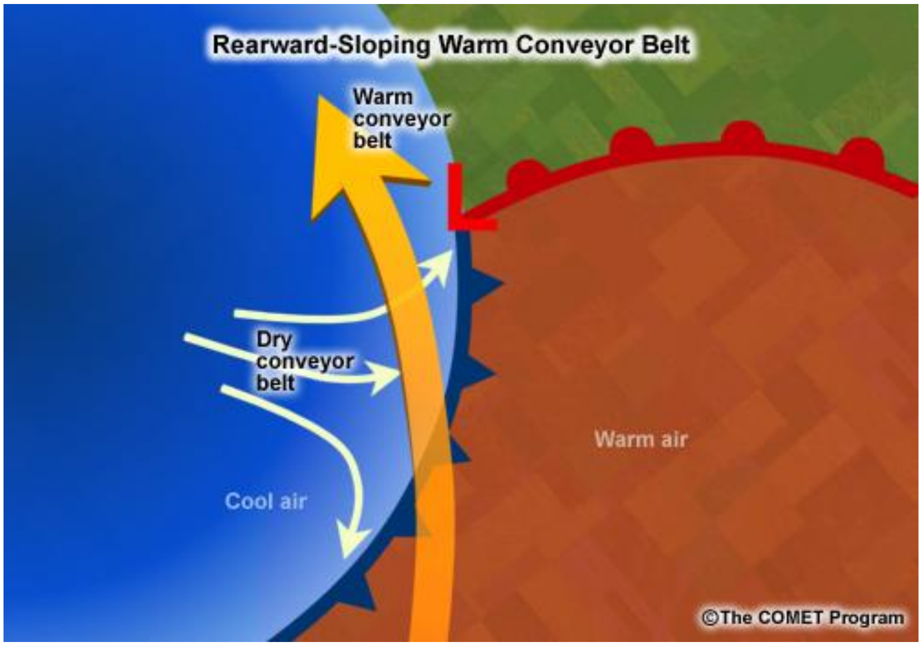


There is a further subdivision of warm conveyor belts into forward-sloping and rearward-sloping ones:

A diffluent upper trough forms a forward-sloping warm conveyor belt that generates precipitation (possibly rainbands) into the warm sector:



A confluent upper trough forms a rearward-sloping conveyor belt that generates strong showers inside the cold front and weaker precipitation in the stratiform clouds behind the front:



12.6. The filling of a low

Cyclolysis: the diminishing of vorticity → disappearance of the low

The vertical axis of geopotential minimum slopes against the flow in developing cyclones. Ascending motion develops downstream of the upper troposphere's vorticity maximum (geopotential minimum), above which one lies divergence.

As cyclogenesis progresses, the bottom of the upper trough gradually approaches the surface low. When the cyclone reaches the occluded stage, the axis of the low rises upright and divergence in the upper troposphere moves downstream of the surface low.

The surface low is at its deepest and surface convergence at its strongest just before the low starts to fill. When there is no more divergence above the surface low, ascending motion ceases and the low begins to fill. Any process that results in lack of divergence above a low pressure system thereby also causes it to fill. This is the case even in the absence of descending motion above the low.

In cyclolysis:

- divergence above the low runs out
- pressure in the low rises
- the low’s vorticity becomes more anticyclonic
- descending motion often develops above the low

12.7. Secondary low

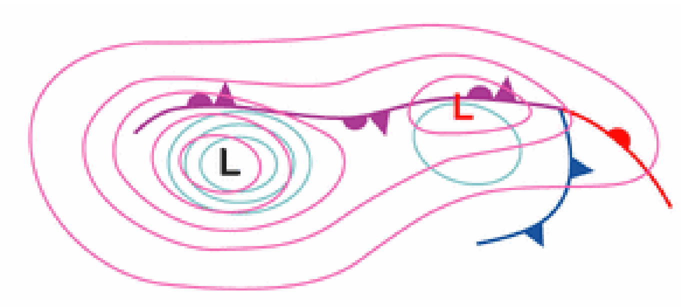
Cyclogenesis does not always stop at the occluded stage. A so-called secondary low, may develop if the following conditions are met:

- both shear and curvature vorticity are strong
- wind velocity in the jet core is at least 50 m/s
- the axis of the jet stream lies on the cyclonic side of the occluded front and is close to parallel to it

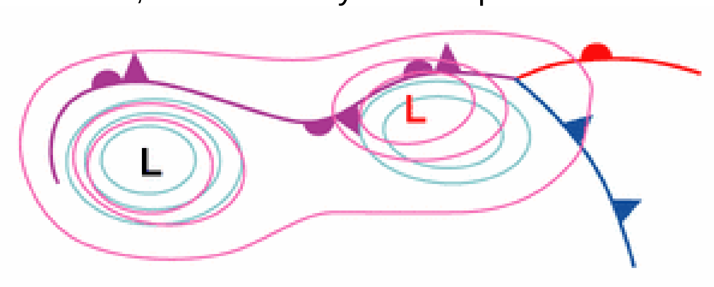
The new low deepens as the old one fills, and the new low will often become deeper than the original.

Frontal analysis of the forming of a secondary low:

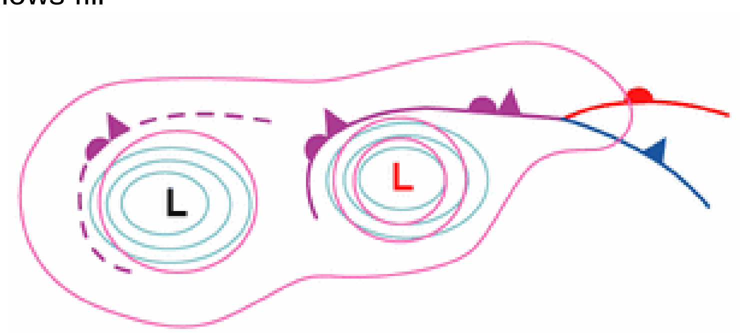
1. the secondary low begins to deepen near the occlusion point when the original low is already far away



2. the original low fills, the secondary one deepens



3. both lows fill



12.8. Potential vorticity

Martin p. 275-295

The original quasi-geostrophic theory examines cyclogenesis using basic parameters such as pressure, temperature and the wind velocity. In an isentropic coordinate system, a similar scrutiny can be carried out with a single variable: PV (potential vorticity, PV perspective).

13.8.1. Conservation of potential vorticity

(Martin, p. 276-277)

As established before, there is a connection between divergence and potential vorticity. The vorticity equation takes the following form in an isentropic coordinate system:

$$\frac{d(\zeta_o + f)}{dt} = -(\zeta_o + f)(\nabla \cdot \overline{V}_o)$$

And the continuity equation takes the following form in the same system:

$$\frac{d}{dt}\left(-\frac{1}{g}\frac{\partial p}{\partial \theta}\right) = -\left(-\frac{1}{g}\frac{\partial p}{\partial \theta}\right)(\nabla \cdot \overline{V}_o)$$

Using the designation

$$\sigma = -\frac{1}{g}\frac{\partial p}{\partial \theta}$$

and through combining and integrating the equations, we get

$$\frac{(\zeta_o + f)}{\sigma} = \frac{(\zeta_o + f)_o}{\sigma_o}$$

This means that the relation

$$\frac{(\zeta_o + f)}{dt} \Big/ \left(-\frac{1}{g}\frac{\partial p}{\partial \theta}\right)$$ is conserved in an adiabatic flow.

In PV perspective this result is formulated as

$$- g(\zeta_\theta + f)\,\frac{\partial\theta}{\partial p} = \text{constant}$$

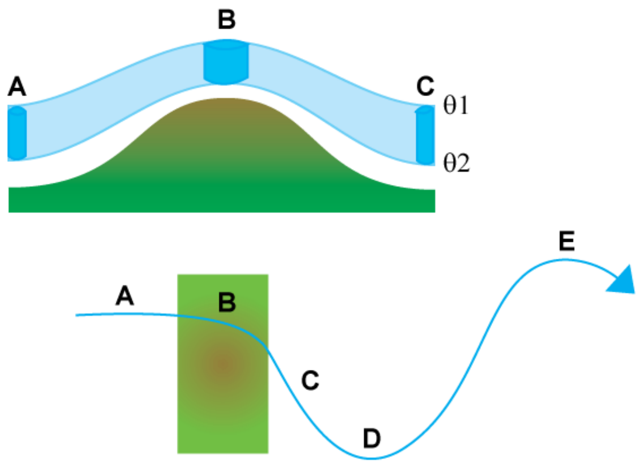
This constant is called isentropic potential vorticity (IPV), but it is commonly shortened to potential vorticity (PV). So, PV is conserved in an adiabatic, homogenous and incompressible flow.

Potential vorticity is a measure of the ratio between an air parcel's vorticity and its effective height. The name comes from an air parcel's potential to increase or reduce vorticity by changing latitudes (f), as well as to increase or reduce stability (the distance between isentropes, $-\partial\theta/\partial p$).

Examples of conservation of potential vorticity:

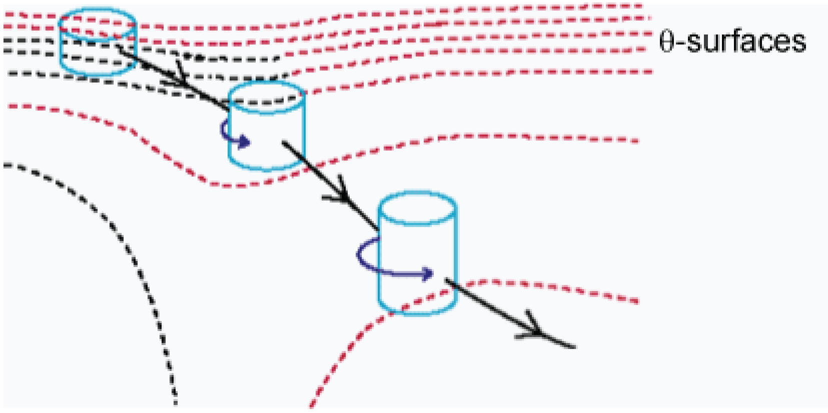
1. An air parcel crosses a mountain range:

   When an air parcel reaches the top of a hill, its height diminishes. This results in a reduction in vorticity, which means vorticity in the parcel becomes more anticyclonic. When an air parcel descends downhill, its height increases and it gains in cyclonic vorticity. In other words, mountain ranges generate rows of waves.



2. air from the lower stratosphere descends into the troposphere:

   when an air parcel from the stratosphere enters the troposphere, its stability is reduced:



12.8.2. Invertibility of potential vorticity

Air flows are not always adiabatic in the real atmosphere; changes in PV are partly caused by variations in friction-induced diabatic heating.

Invertibility

From the distribution of potential vorticity one can, in certain conditions, get:

- vorticity (horizontal winds)
- static stability (vertical distribution of temperature)
- geopotential
- ageostrophic wind

The prerequisitions for getting these parameters are

- distribution of PV in a given location
- conditions near the borders of the area
- the balance between the mass and momentum of air

→

PV inversion

12.8.3. Potential vorticity and release of latent heat

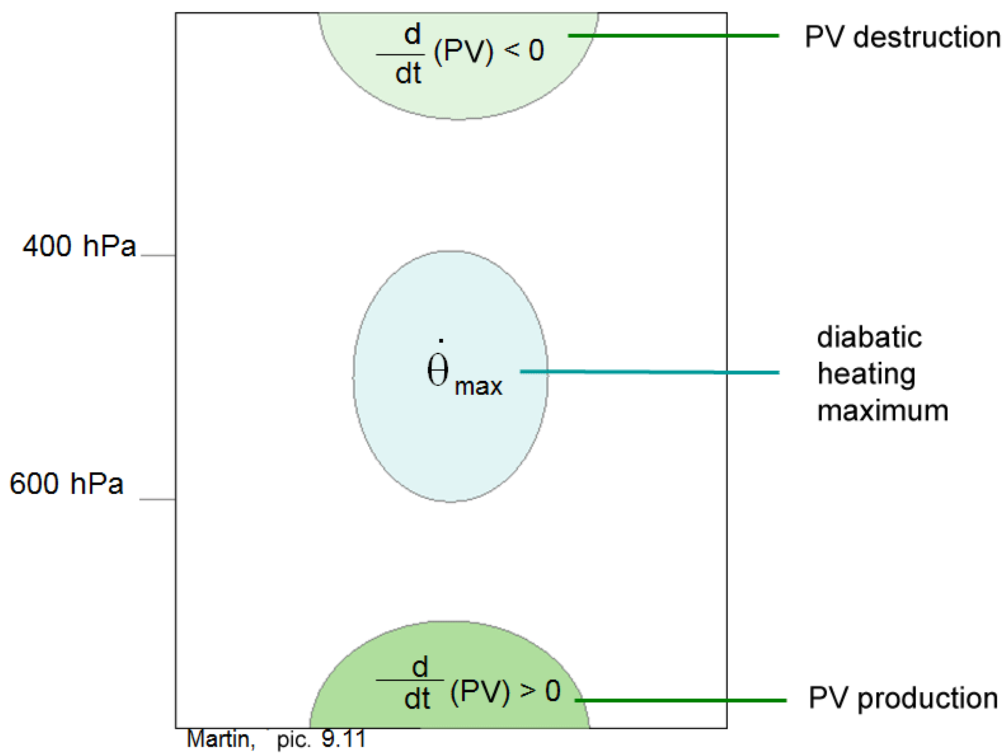
The three-dimensional potential vorticity equation (Ertel 1942):

$$\frac{d(PV)}{dt} \approx -g(\zeta + f)\frac{\partial\dot{\theta}}{\partial p}$$

where theta point = rate of diabatic heating.

In other words: potential vorticity increases when the vertical gradient of diabatic heating is positive, and decreases when the gradient is negative.

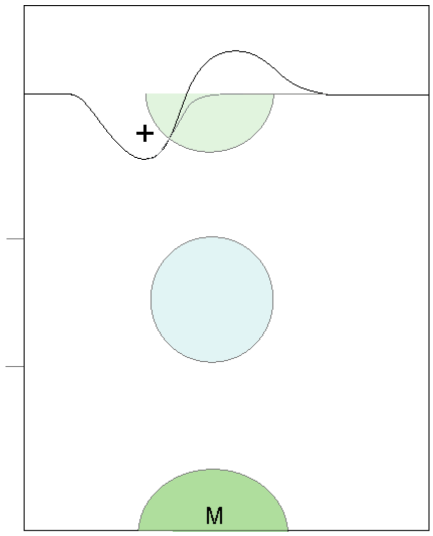
In a typical mid-latitude cyclone, diabatic heating is strongest in the mid-troposphere:



The maximum of heating and ascending motion lies slightly ahead of the positive PV anomaly above it. This causes the PV anomaly to strenghten in the lower troposphere and to weaken high up.

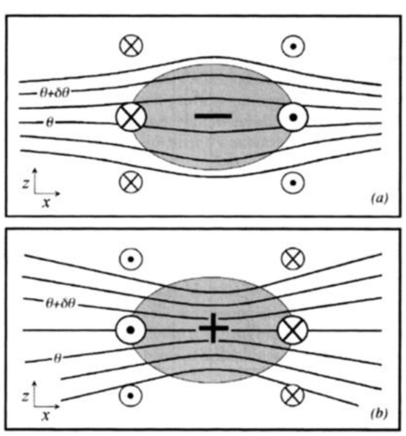
The maximum of heating and ascending motion lies slightly ahead of the positive PV anomaly above it. This causes the PV anomaly to strenghten in the lower troposphere and to weaken high up.

Meanwhile, the PV anomaly down below acts together with the one high up to strengthen them, in the process also intensifying the surface low:



12.8.4. Potential vorticity anomalies

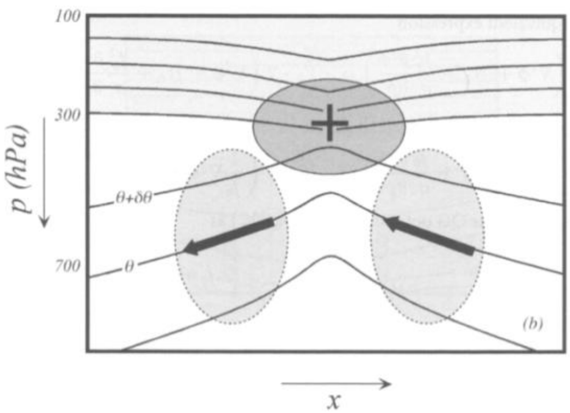
For the purpose of dynamical examinations, PV anomalies with either a local maximum or minimum are especially useful. Compared to the environment, isentropes are spread further apart in a minimum and closer together in a maximum. If the anomaly is located in the tropopause, wind speed also reaches its maximum there:



(Martin, figure 9.5)

When the invertibility of potential vorticity is applied in dynamics, PV maximums (positive PV anomalies) are a subject of particular interest. Anomalies of this kind have local maximums of both vorticity and static stability.

A positive PV anomaly in the tropopause:



(Martin, figure 9.6.)

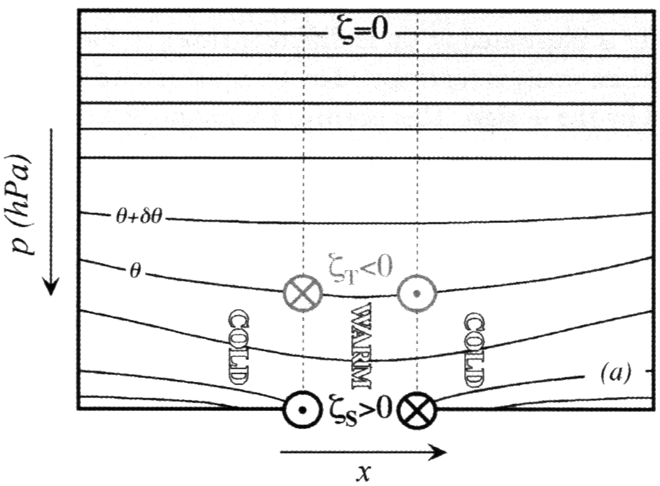
The tropopause lies at 300 hPa in the figure above, and there is a positive PV anomaly traveling along it to the right. θ(e)– surfaces are closest to one another near the anomaly as well as above and below it. As the anomaly moves, an adiabatic flow moves along the isentropes to the left (black arrows, the so-called vacuum cleaner effect). In these conditions there is ascending motion downstream and descending motion upstream of the anomaly.

This is identical to the previous observation: there is ascending motion ahead of cyclonic vorticity maxima and descending motion behind them.

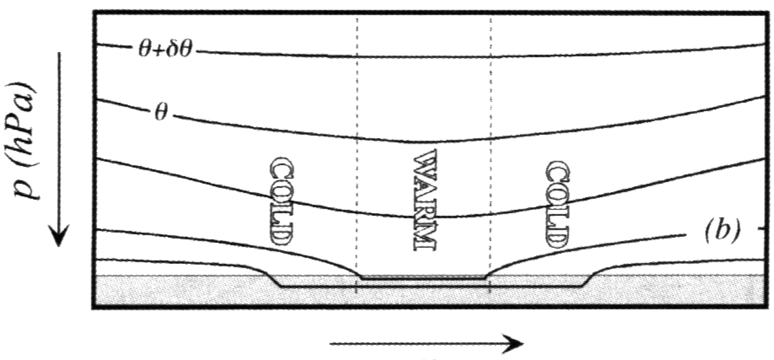
The influence of the PV stretches to a certain height both upwards and downwards of the maximum. This height is called the penetration depth H = fL/N, where L = the anomaly's horizontal lenght, and N = the Brunt-Vaisala frequency.

Anomalies on the surface are also important for cyclogenesis. Baroclinic zones in the lower troposphere and the temperature anomalies associated with them can also be considered PV anomalies. The following figure depicts the warmest anomaly of potential temperature on the surface (which could also be the warm sector of a cyclone).

There is no wind in the tropopause, which means there is zero vorticity there. The air above a warm anomaly on the surface is warmer than its environment, which generates thermal anticyclonic vorticity in the mid-troposphere, driving the isotherms further apart. At the surface there is cyclonic vorticity, and the isotherms are drawing closer to each other (figure 9.7.a), because the total vorticity has to be zero. The process of isotherms approaching one another can be made clearer by drawing the sections under the surface (figure 9.7.b).



(Martin, figure 9.7.a)



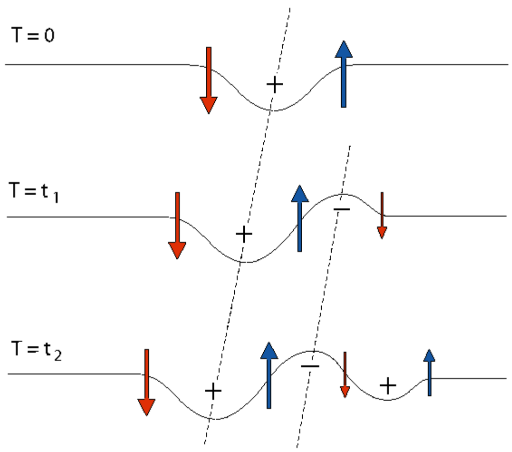
(Martin, figure 9.7.b)

In summary, warm anomalies near the surface are positive PV anomalies.

12.8.5. Cyclogenesis from the PV perspective

Let us first consider PV anomalies in the upper and lower troposphere separately.

The following figure shows the development over time of a positive PV anomaly in the tropopause in a westerly flow.

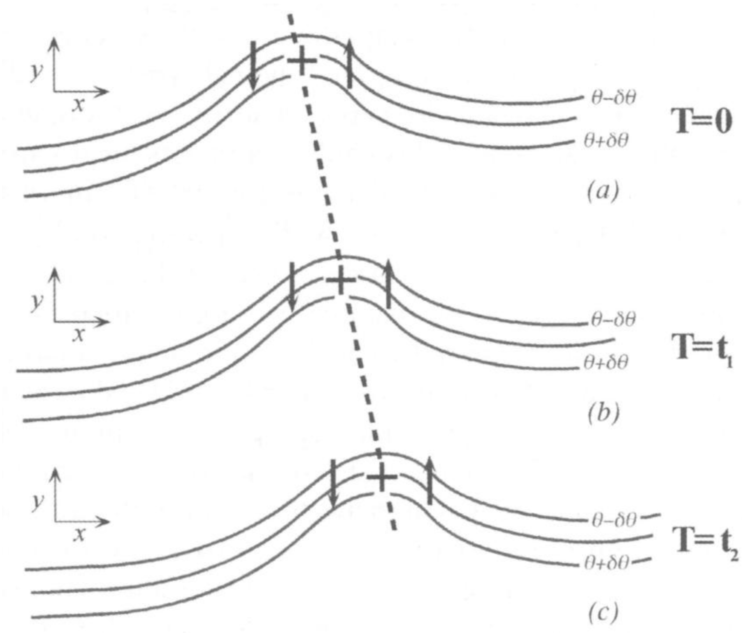


Circulation of the anomaly is marked in red (cyclonic) and blue (anticyclonic) arrows. At T = 0 the circulation is cyclonic. There is anticyclonic advection to the east of the anomaly and cyclonic advection to the west of it. As a consequence, the conditions at T = 1 are such that a small, anticyclonic anomaly has developed east of the original anomaly, while the original has moved west (left in the figure). At T = 2, the original anomaly has continued westward, and a new, small cyclonic anomaly has formed to the east of anticyclonic anomaly.

If an anomaly like this gets to develop outside the influence of other factors, it will travel upstream (just like large-scale Rossby waves and retrogression of upper waves).

Positive PV anomalies near the surface

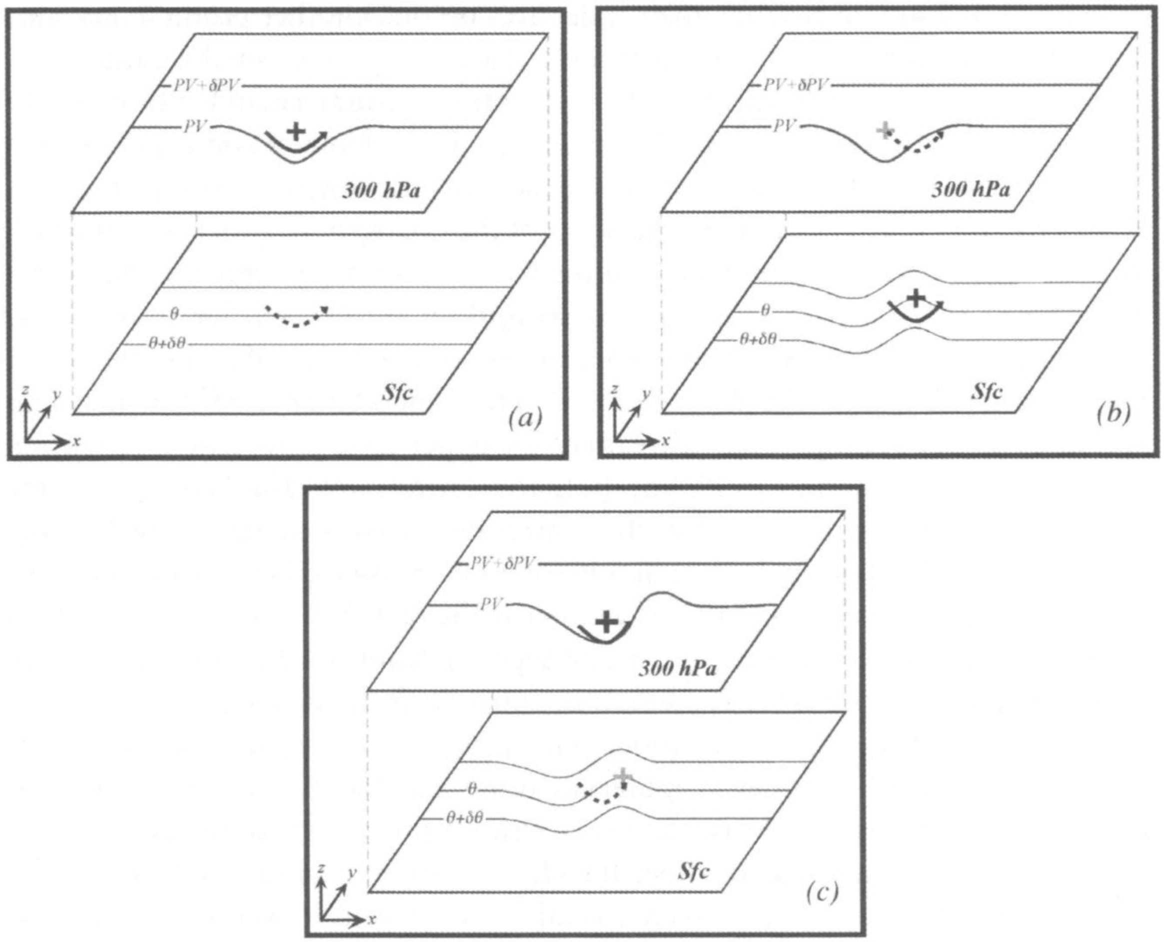
> When there is a positive anomaly on the surface, cyclonic circulation will generate a warm anomaly there. The anomaly will move east in a westerly flow, and this being the case, southerly winds downstream of the anomaly will bring warm advection there. Conversely, cold air is advected upstream.



(Martin, figure 9.9)

Disturbances in both the upper and lower troposphere have an influence that also extends vertically. If their relative positions are favorable, anomalies in the tropopause and at the surface may intensify each other. Extended periods of such mutual influence result in cyclogenesis.

Cyclonic circulation in the upper troposphere reaches the surface (though faintly), where it alters the isotherm field and generates warm advection. The warm advection forms a warm anomaly, which is also the positive PV anomaly of the surface. This anomaly has its own circulation with its own upward influence; it intensifies the PV anomaly above it by supplying its eastern parts with positive PV advection. On the other hand, there is negative PV advection to the east of it, and the combined effect of the two advective forces is the anomaly's continued movement along the flow.



a) an anomaly in the upper troposphere influences the lower anomaly

b) circulation in the lower troposphere influences the upper anomaly

c) circulation in the upper troposphere intensifies and its reflection down below strengthens the temperature advection

The PV anomaliy in the upper troposphere is located upstream of the pressure anomaly. Due to its influence, warm advection is strongest in the middle of the warm anomaly, and cold advection is strongest upstream of it. This causes the warm anomaly in the lower troposphere to intensify and, against its tendency, to move against the flow.

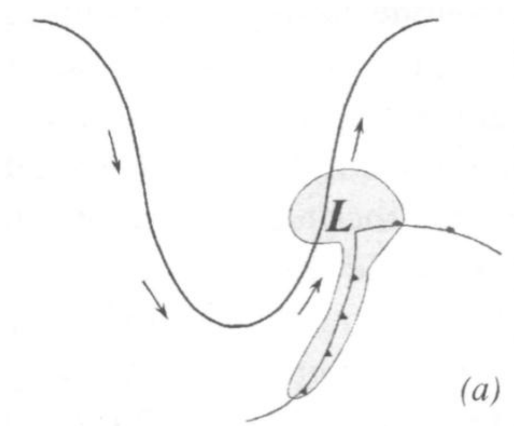
That is to say: anomalies in the upper and mid-troposphere oppose each other's movements, which allows them to interact long enough for cyclogenesis to initiate and intensify.

Also from this perspective cyclogenesis can only begin when a cyclone's axis tilts upstream with height.

12.8.6. Occlusions from the point of view of PV   Martin p. 297-302

The occlusion process can be described with a PV field of the tropopause in the following way:

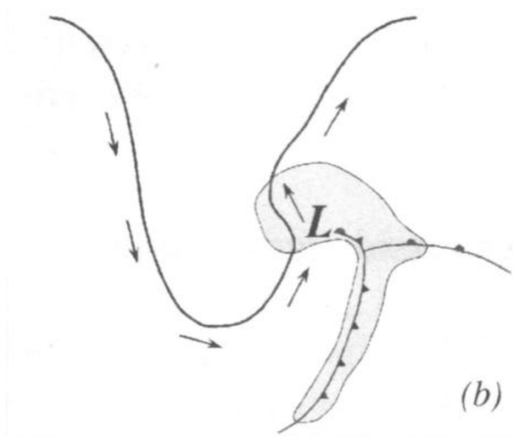
1. Developing cyclone



(Martin, figure 9.16)

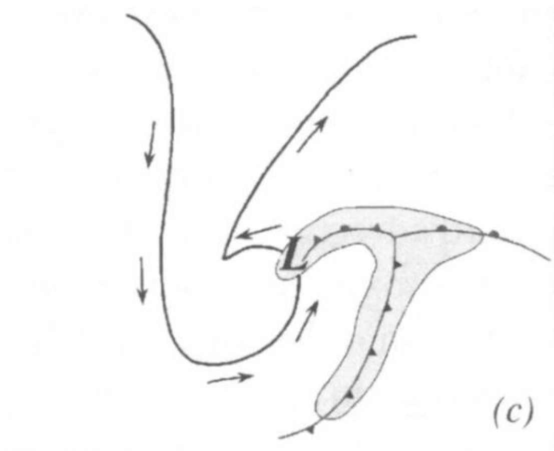
The black isoline represents the tropopause as PV value; for example, PV = 2 PVU.
The arrows denote an air flow in the tropopause, and the shaded area contains PV loss caused by diabatic heating. L is the location of the surface low that the pictured surface fronts are part of.

2. Beginning of an occlusion



The destruction of PV alters the PVU field and tropopause air flow around the developing occluded front.

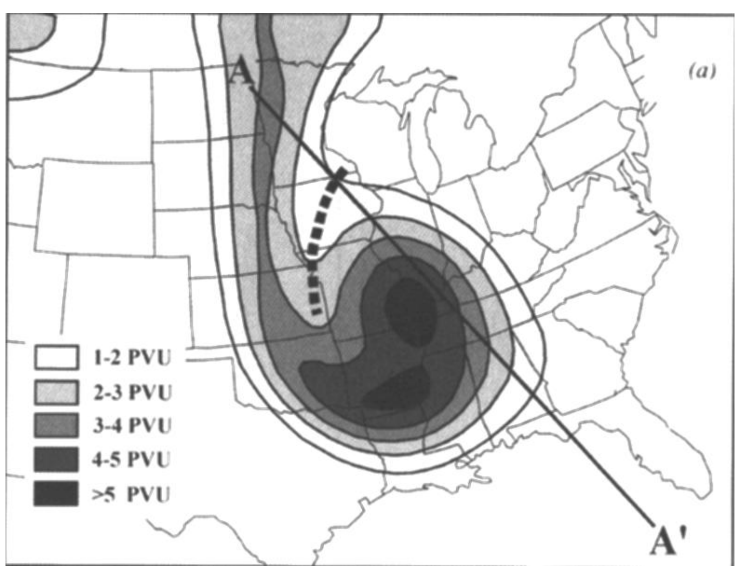
3. Old occlusion



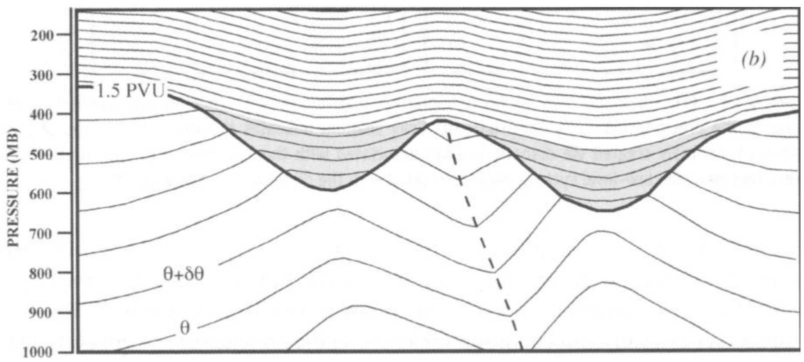
The center of the surface low is far from the peak of the warm sector, and the region of diabatic heating has moved further away from the notch that has developed in the PVU field. Negative PV advection causes the notch to intensify. Finally the development of the PV notch brings about the detachment of the PV maximum from the rest of the anomaly, and a separate warm layer forms below it.

If we look at the temperature across the whole troposphere, the warm layer lies to the northwest of the warm ridge near the surface. (In the figure, north is up and east to the right.)

On a larger scale, the situation could look like this:



The vertical cross section:



This is similar to the structure found in warm occlusions.

13. The tropopause

13.1. Dynamic tropopause

The tropopause is an atmospheric zone where the lapse rate and static stability change. The tropopause separates the very stable stratosphere from the less stable troposphere.

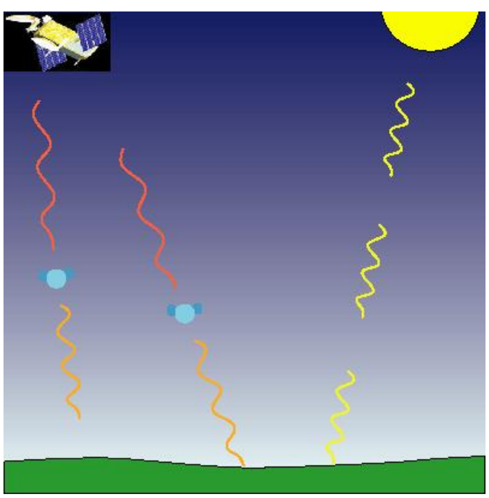
Because potential vorticity is the product of vorticity and stability, it can be used to determine the height of the tropopause. When defined in this manner, the tropopause can be called by the more specific name of dynamic tropopause.

PVU (potential vorticity unit): $\dfrac{10^{-6}\,K}{s\,kg\,m^{2}}$

PVU is > 3 in the stratosphere and < 3 in the troposphere. In mid-latitudes the tropopause is mostly defined as 2 PVU, but also values 1 and 1,5 are used.

13.1.1. A topographic representation of water vapor image's brightness temperature

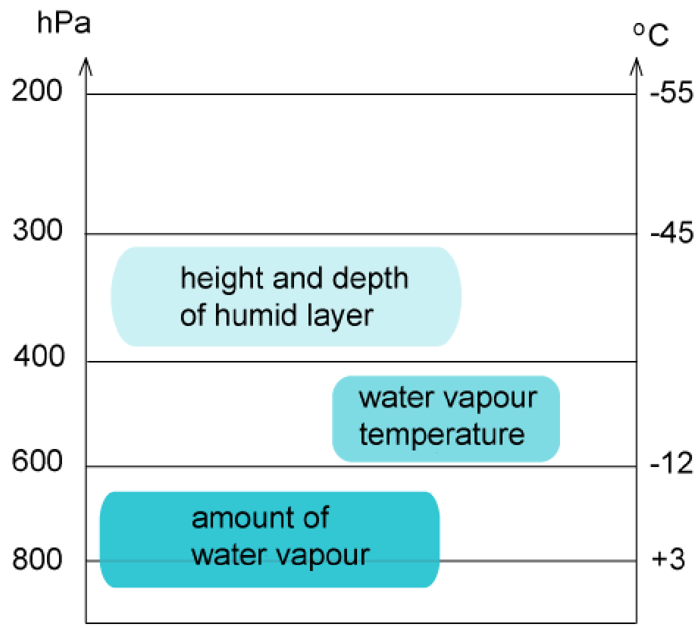
In water vapor channels (5-7 µm) the weather satellites receive the thermal radiation emitted from water molecules:



The wavelength of the thermal radiation satellites receive depends on

- water molecule quantity
- water molecule temperature
- thickness and height of humid layers

Typically, the higher up a humid layer is, the colder it is and the less water it contains.

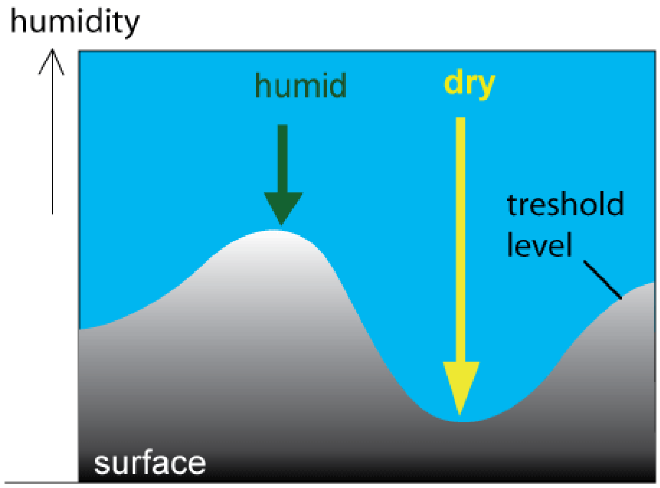


The topographic representation is a simplified model of the greyscale of water vapor channels. In the troposphere there is a height called the threshold level, below which the air is humid enough for water vapor to absorb all radiation coming from the surface. Satellites cannot see below this level with water vapor channels.

The assumptions of the topographic representation:

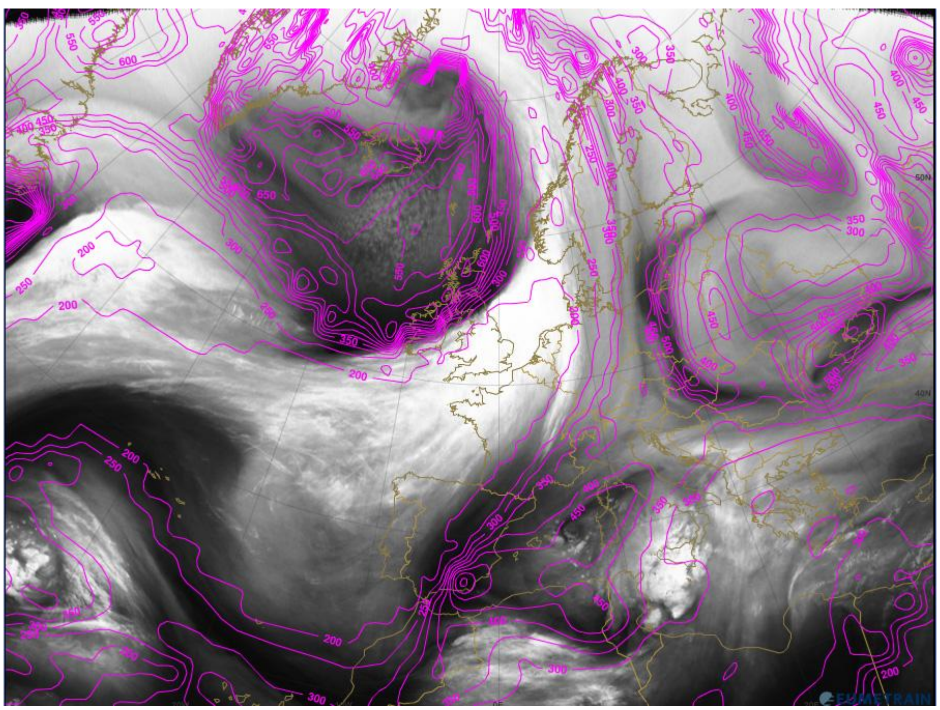
- lower troposphere is humid up to the threshold level
- air above the threshold level is relatively dry and there are no moist layers
- air cools with height and there are no significant inversions
- no clouds present

When the threshold level rises, brightness temperature falls and the gray shade turns lighter. Water vapor imagery can be seen as a topography of humidity called a moisture terrain:



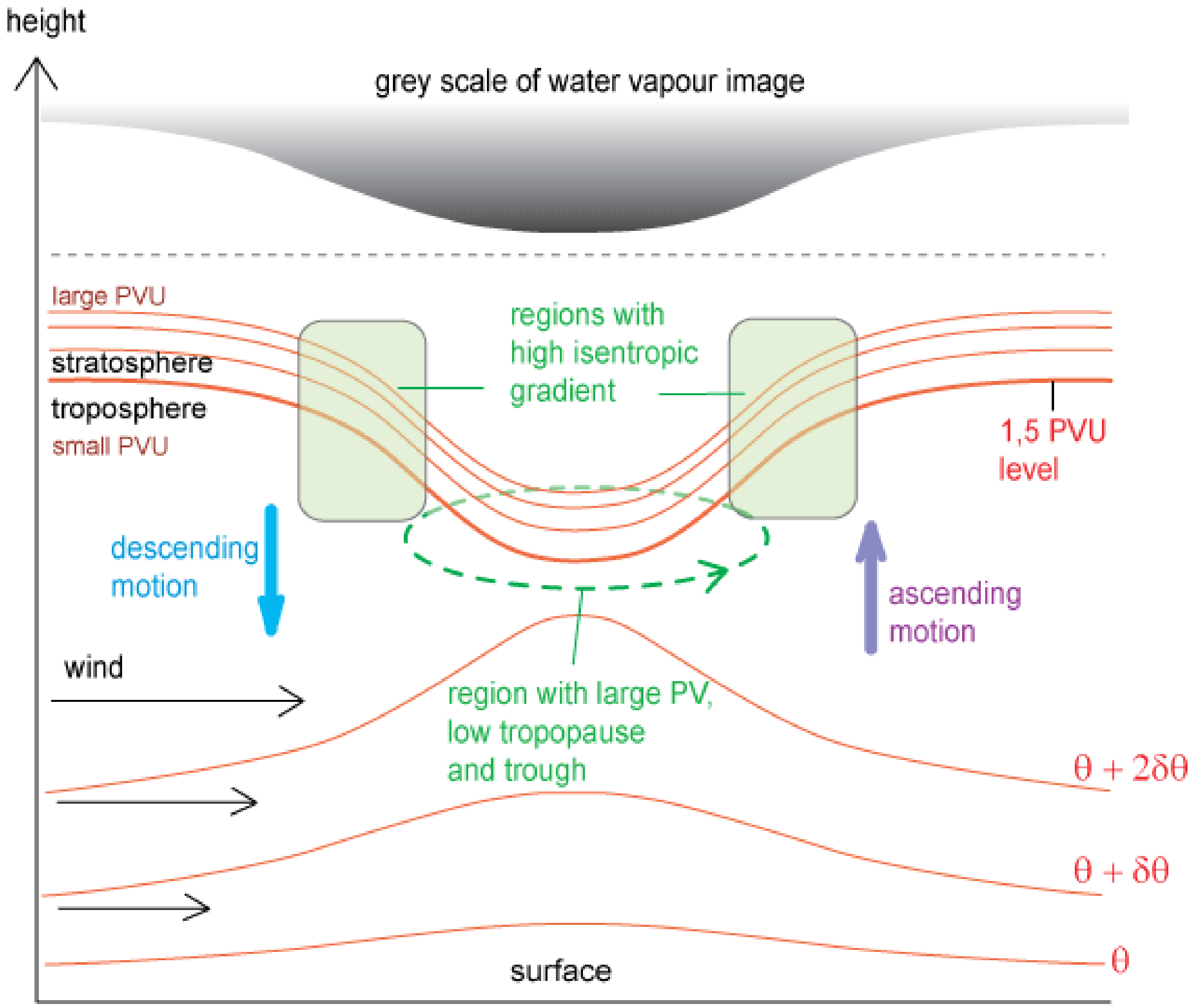
The height of the threshold level is dependent on temperature: because warm air can hold more moisture than cold air, the threshold lies lower in cold areas. The first assumption is usually true for the real atmosphere. That being the case, interpretation of the imagery depends on inversions, clouds and moisture layers.

The height of the tropopause also indicates the altitude of jet streams. This ePort image shows a numerical field that gives the height of the PVU=1 surface, interposed over the 6.2 µm water vapor channel:



The properties of a positive PV-anomaly moving in the tropopause:

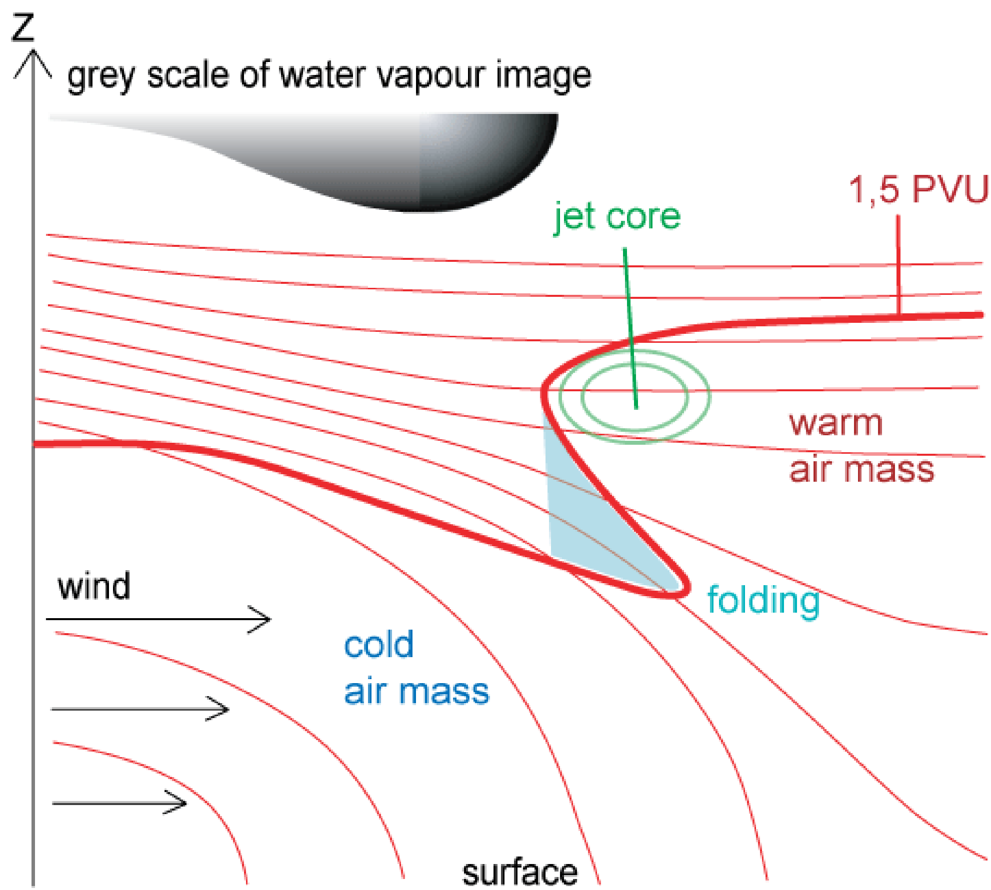
- separates the troposphere from the stratosphere
- local PVU maximum
- local tropopause height minimum
- local geopotential minimum
- ascending motion ahead, descending motion behind
- air below is more labile than in its surroundings



13.2. Tropopause folding

Tropopause folding means the adiabatic descent of stratospheric air into the upper troposphere along the isentropes. This has effects on both the climate at large as well as air quality in the troposphere: harmful substances such as ozone, methane, nitrous oxide and CFCs come from the stratosphere.

Folds reach down to depths of 1-4 km. Folding is related to jet streams and fronts (high vertical shear and tight temperature gradient). More on this in chapter 14.



14. Fronts

Martin p. 189-192
Lackmann p. 131-163

14.1. Frontogenesis and frontolysis

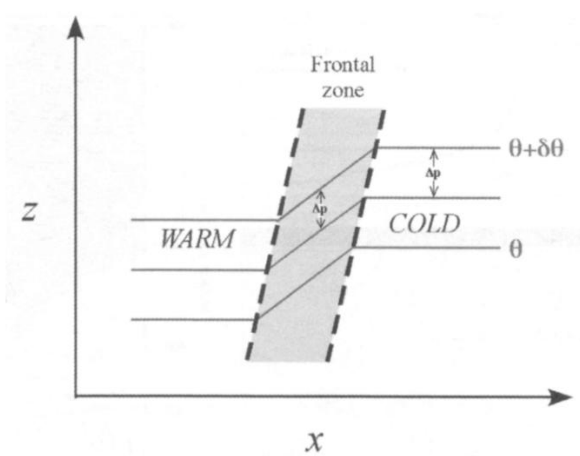
14.1.1. Mid-latitude frontal structure

What exactly is a front? Martin defines a front as follows: a boundary whose primary structural and dynamical characteristic is the larger-than-background temperature (or density) contrast associated with it.

Other definitions:

- a cold or warm front is a zone in which the temperature changes more when traveling through it than it does traveling through its environment.
- an air mass boundar
- a baroclinic boundary

In the Norwegian model, fronts are regions of sharp discontinuity in temperature, also known as zero-order fronts. However, actual fronts would be better characterized as first-order fronts, since the discontinuous parameter in them is temperature gradient rather than temperature itself:



Martin, figure 7.3.

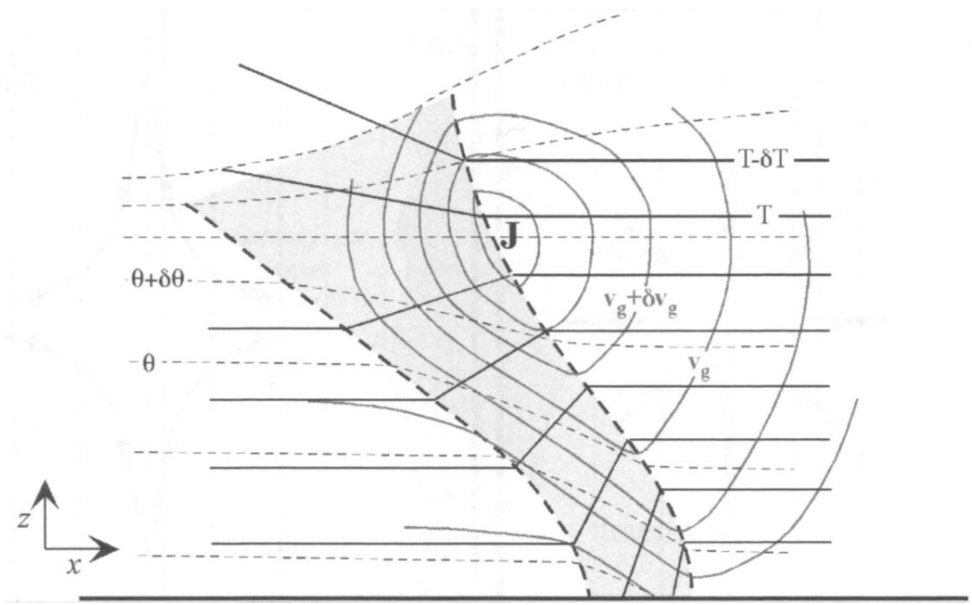
In the end, however, the zero-order approximation presents the simplest way of studying the properties of fronts. Assuming that no parameter changes in the direction of the front, that the pressure tendency is zero and that hydrostatic and geostrophic equilibria are in effect, the end result indicates the presence of positive relative vorticity in fronts.

Fronts have

- horizontal temperature gradients stronger than their environment
- stronger relative vorticity than their environment
- more static stability than their environment
- strong vertical shear

14.1.2. Frontogenesis and vertical motion

A vertical cross section of a frontal zone:



Martin, figure 7.5

Thermal wind law:

$$\overline{V}_T = \overline{V}_g(p) - \overline{V}_g(p_o) = \frac{R}{f}\ \overline{k} \times \nabla_p \overline{T}\ \ln\frac{p_o}{p}$$

$\overline{T}$ = mean temperature of the layer $(p_o - p)$

The temperature gradient is strongest near the surface and weakens with height.

Because fronts have strong temperature gradients, the thermal winds are likewise strong. This means the vertical shear of geostrophic wind is also strong. Wind speed increases with height and reaches its maximum at the tropopause, where it forms the core of a jet stream.

The ’location of a front’ is the leading edge of the frontal zone. Horizontal winds are stronger inside a front than in its surroundings. Because in a (s,n) coordinate system the geostrophic vorticity is $\zeta = \frac{V}{R_s} - \frac{\partial V}{\partial n}$,

the front must also contain a local vorticity maximum. This being the case, there must be convergence inside the front and divergence above it, which means fronts contain ascending motion.

A front will have local horizontal maxima for the following parameters:

- temperature gradient
- vorticity
- wind velocity
- convergence
- ascent

If the temperature gradient tightens, the thermal winds become stronger, as does the wind at the core of the jet stream. This, in turn, increases vorticity, causing convergence and ascending motion to increase and the whole front to strenghten.

14.1.3. The frontogenesis function

Lackmann p.135-141

Frontogenesis is a process that increases the (potential) temperature gradient, diminishing the distance between isotherms.

In particular, horizontal frontogenesis is a horizontal advective process in which the temperature gradient tightens. Generally the frontogenesis function F is

$$F = \frac{D|\nabla\theta_p|}{Dt}\text{, where}$$

$$\frac{D}{Dt} = \frac{\partial}{\partial t} + \bar{v}\cdot\nabla_p + \omega\frac{\partial}{\partial p}$$

(Pettersen 1936)

This function is defined in a Lagrangian coordinate system, which is to say: as a change in the potential temperature gradient of the air traveling with a flow. It is better to define frontogenesis using potential temperature rather than temperature, because doing so reduces the effect that orography has on the process.

Frontogenesis generates and intensifies frontal zones, while frontolysis erodes them.

The three-dimensional form of the frontogenesis equation contains over a dozen terms. For the sake of simplicity, let us consider the change in temperature gradient in the direction perpendicular to the frontal zone, y.

When x is be parallel to the frontal zone, the frontogenesis function is

$$F = d/dt\,(\partial\theta/\partial y)$$

In component form:

$$F = \frac{\partial\theta}{\partial x}\frac{\partial u}{\partial y} + \frac{\partial\theta}{\partial y}\frac{\partial v}{\partial y} + \frac{\partial\theta}{\partial p}\frac{\partial\omega}{\partial y} - \frac{\partial}{\partial y}\left(\frac{\partial\theta}{\partial t}\right)$$

These four terms correspond to the four physical processes that affect the change of the potential temperature gradient:

- shearing
- confluence
- tilting
- diabatic heating

1. Shear $\frac{\partial \theta}{\partial x}\frac{\partial u}{\partial y}$

The shear term describes changes that are caused by the wind component parallel to the front.

The figure below offers an idealized representation of shear-induced frontogenesis in a cold front.

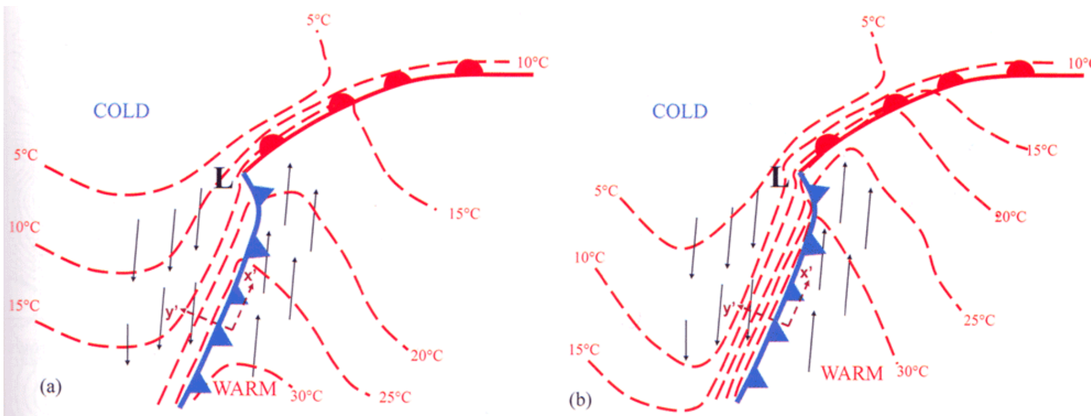


Figure 6.3. Idealized horizontal plot of near-surface wind vectors (black arrows), isotherms (red dashed contours), and frontal zones for (a) initial time, and (b) the same front-relative view at a later time (~24 h later). The rotated front-relative coordinate axes are shown.

In part (a) $\partial\theta$ weakens in the direction of xand the wind component u parallel to x weakens in the direction of y, which means both components are negative and the shear term itself is positive. This implies that frontogenesis is taking place, which means the temperature gradient is tightening (and the cold front intensifying), as shown in part (b).

In the same scenario, the warm front is weakening (undergoing frontolysis):

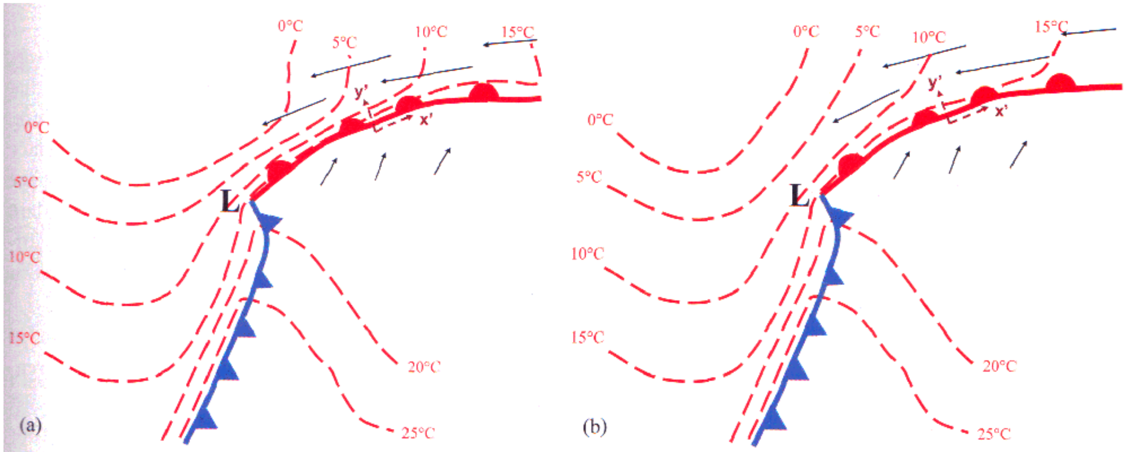


Figure 6.4. As in Fig. 6.3, but emphasizing the shearing process in the warm-frontal region.

In part (a) u weakens in the direction of y, but to the north of the front θ increases in the direction of x, so the shear term is negative.

2. Confluence $\dfrac{\partial \theta}{\partial y}\dfrac{\partial v}{\partial y}$

The confluence term describes changes caused by the wind component perpendicular to the frontal zone. It is larger than the shear term by an order of magnitude.

The confluence term can also be called the stretching term.

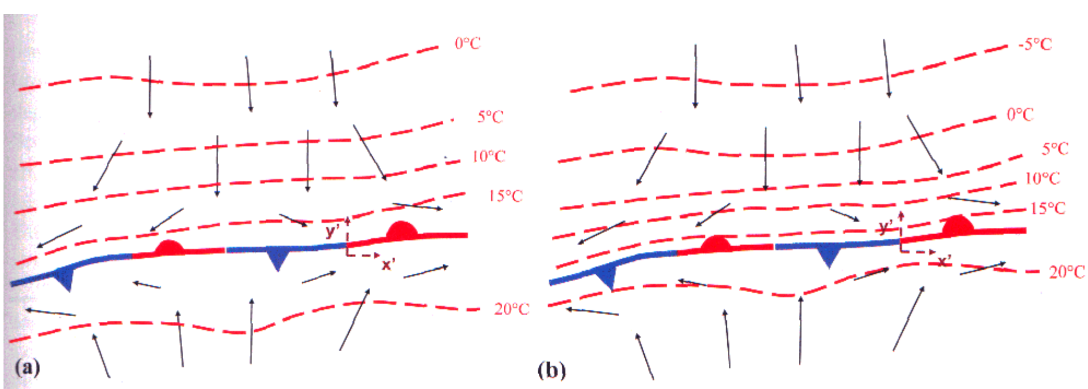


Figure 6.5. As in Fig. 6.3, but demonstrating confluent frontogenesis.

In the scenario above, the warm side of a stationary front is warming and the cold side is cooling, which means the front is intensifying. θ weakens in the direction of y (turns negative when crossing the frontal zone). The product of the components is positive, which means frontogenesis is taking place.

The confluence term can also generate frontolysis, depending on the angle between the flow and the isotherms.

3. Tilting $\dfrac{\partial\theta}{\partial p}\dfrac{\partial\omega}{\partial y}$

Vertical motions are typically relatively weak near the surface, so the effects of the tilting term are small. Usually it generates frontogenesis in the mid- and upper troposphere.

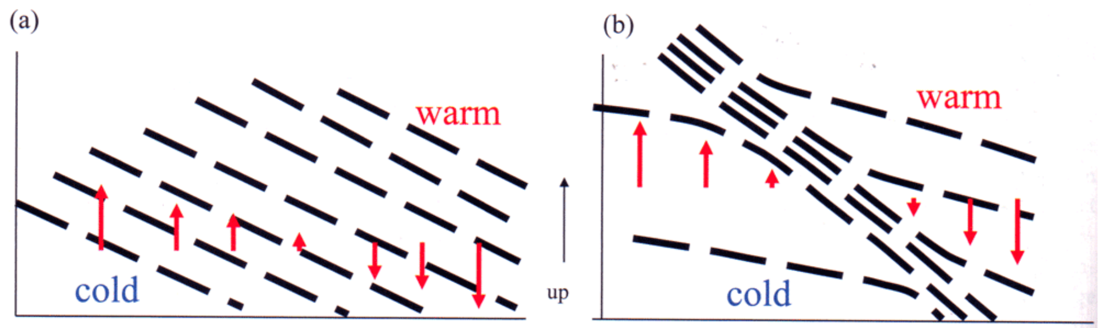


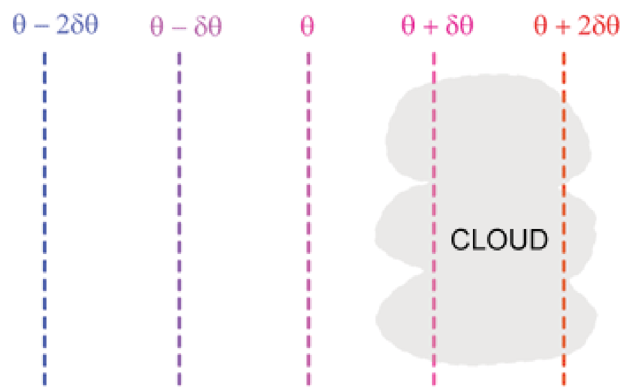
Figure 6.6. Idealized cross-sectional diagrams demonstrating frontogenetical tilting. Dashed lines are isentropes and red arrows represent vertical motion for (a) an initial time and (b) a later time (after Carlson 1998).

The situation depicted above could take place on the left exit region of a jet stream, where there is thermally indirect vertical circulation. Differential temperature advection intensifies the front.

Often the tilting term, tending to maintain the thermal wind equilibrium, acts against the shear and confluence terms.

4. Diabatic heating $\frac{\partial}{\partial y}\left(\frac{\partial \theta}{\partial t}\right)$

This factor relies on environmental conditions such as cloudiness, season and time of day. The following example is for a situation in summer with a baroclinic zone where the warm side is cloudy, and the cold side cloud-free:



The sun warms the cold area more than the cloudy one during the day, so the temperature gradient weakens over time; frontolysis is taking place. At night, on the other hand, the the cold side cools faster than the warm side, which means there is frontogenesis in the area.

Large horizontal temperature and moisture gradients develop easily in relatively small locations with sharp changes in their surface properties. For example:

- snow-covered / snow-free ground
- frozen / ice-free sea
- cold sea / warm ground or warm sea / cold ground

In these conditions very low level baroclinic zones are formed. They may strengthen cyclones beyond what might be expected of lower and mid-troposphere baroclinity alone.

The two-dimensional frontogenesis function

The previous section presented the frontogenesis terms of one direction only. Frontogenesis can also be examined in simplified form on a two-dimensional x-y plane. Examination of this sort is suited for most fronts, since tilting terms are relatively small.

$$F_{2D} = \frac{|\nabla\theta|}{2}(F\cos 2\beta - D)$$

Here F = total deformation, D = divergence, and (symb) = the angle between isotherms and the axis of dilation.

Frontogenesis can therefore be the result of

1. convergence

2. total deformation influencing the isentropes so that the angle between its axis of dilation and the isentropes is 0 – 45°.

The effect of deformation on the distance between isentropes in two extreme cases:

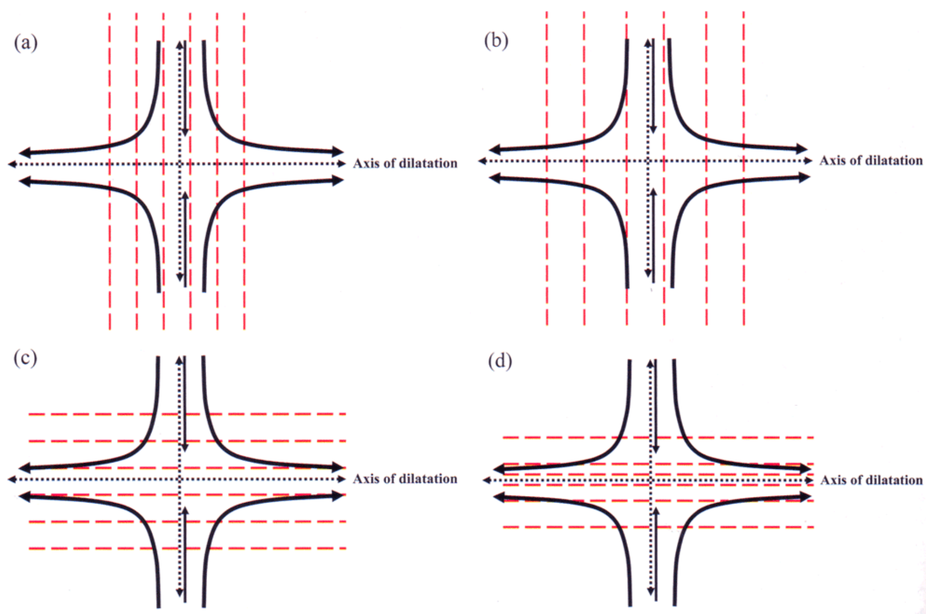
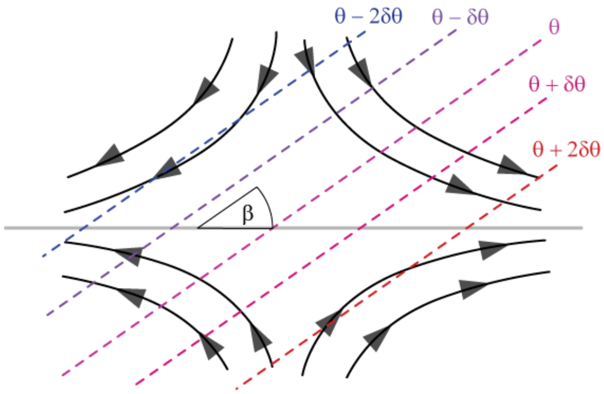


Figure 6.9. Idealized schematics of isentrope evolution in a deformation flow. Black lines are streamlines, and red dashed lines are isentropes: (a) initial time, isentropes oriented perpendicular to axis of dilatation; (b) as in (a), but at a later time. (c),(d) As in (a),(b), but with isentropes oriented parallel to the axis of dilatation.

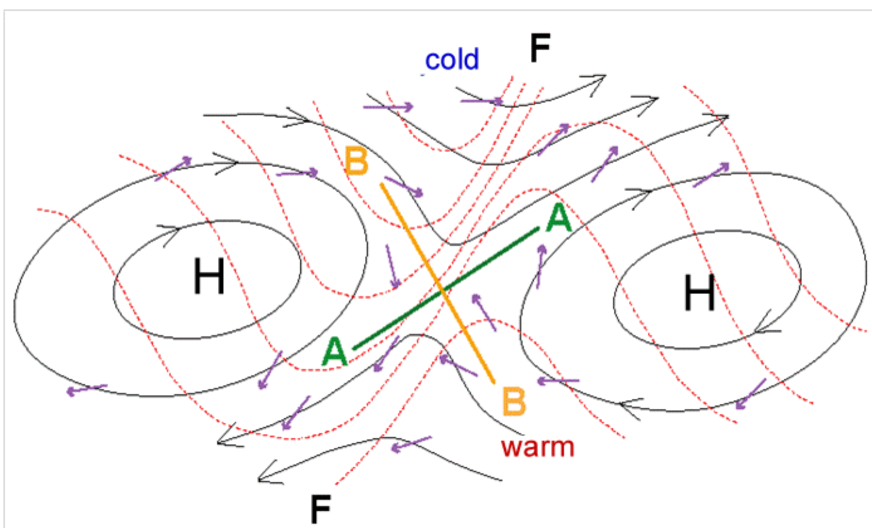
In (a) the isentropes are perpendicular to the axis of dilation, resulting in frontolysis, as seen in (b). In (c) the isentropes are parallel to the axis of dilation, which results in frontogenesis, as seen in (d).

More generally:



Deformation fields favorable to frontogenesis can be found around cols, for example.

Palmen, Newton 1969: Frontogenesis on the col F - F, where A - A is the axis of dilation and B - B the axis of stretching. The geostrophic flow is marked in gray and the friction-induced boundary layer wind in purple:



Many numerical models give frontogenesis calculated with the shear and confluence terms.

Fronts accompanying deepening lows and troughs are stronger than those accompanying intensifying highs and ridges. When there is a high on the surface, it may happen that the surface front cannot be found at all, even if a frontal zone is clearly present above.

14.1.4. The effects of ageostrophic wind

In frontal zones, the geostrophic assumption is largely valid in the direction of the frontal zone, where the ageostrophic component is small compared to the geostrophic component. On the other hand, geostrophy does not apply for the direction perpendicular to the frontal zone.

Assuming that the ageostrophic component aligned with the frontal zone is much smaller than the geostrophic one, the ageostrophic part can be studied with the semi-geostrophic Sawyer-Eliassen equations. Leaving out the tilting terms, the geostrophic forcing Q(g) of frontogenesis can be formulated on a x-y plane as:

$$Q_g = -2\left(\frac{\partial U_g}{\partial y}\,\frac{\partial V_g}{\partial p} - \frac{\partial V_g}{\partial y}\,\frac{\partial U_g}{\partial p}\right)$$

y in the previous equation is the ageostrophic component of a flow perpendicular to the frontal zone. Using the thermal wind law, the equation can be written as

$$Q_g = 2\gamma\left(\frac{\partial U_g}{\partial y}\,\frac{\partial \theta}{\partial x} + \frac{\partial V_g}{\partial y}\,\frac{\partial \theta}{\partial y}\right)$$

Here γ is a coefficient defined according to the hydrostatic assumption. It only depends on pressure.

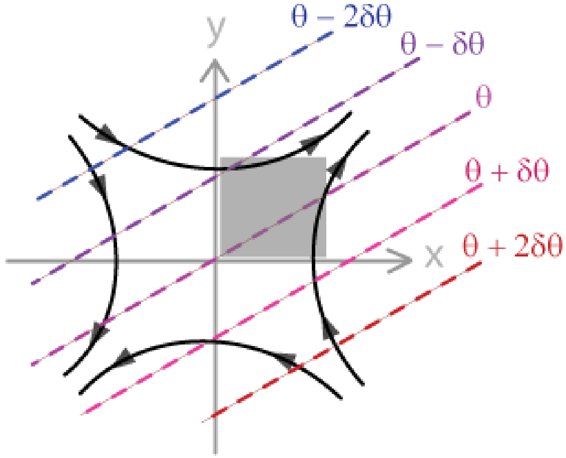
The first term on the right side of the equation is geostrophic shear defomation, and the second one is geostrophic stretching deformation.

When tilting terms are taken into account, the effects of the ageostrophic wind component are simplest to describe using Q-vectors.

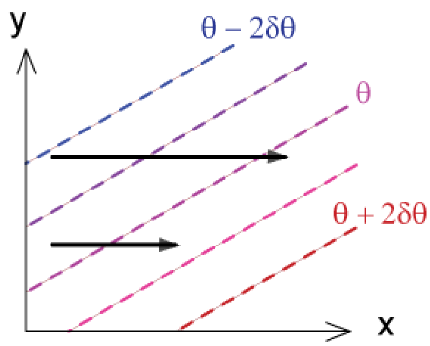
An example of shear deformation:

$$Q_{g_{SH}} = 2\gamma\,\frac{\partial U_g}{\partial y}\,\frac{\partial \theta}{\partial x}$$

Frontogenetic deformation field:



a close-up of the gray box:

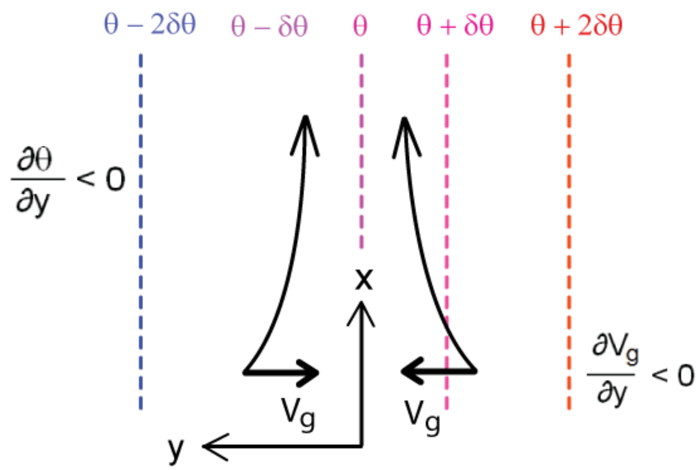


In this case

$$\frac{\partial U_g}{\partial y} > 0 \quad \text{and} \quad \frac{\partial \theta}{\partial x} > 0 \text{, so} \quad Q_{g_{SH}} > 0$$

An example of stretching deformation:

$$Q_{g_{ST}} = 2\gamma \frac{\partial V_g}{\partial y} \frac{\partial \theta}{\partial y}$$



Both terms are negative in this confluent flow, which means stretching deformation is positive.

The solutions to the Sawyer-Eliassen equations imply that frontogenesis is a two-stage process:

1. geostrophic deformation tightens the temperature gradient, generating a secondary ageostrophic indirect circulation

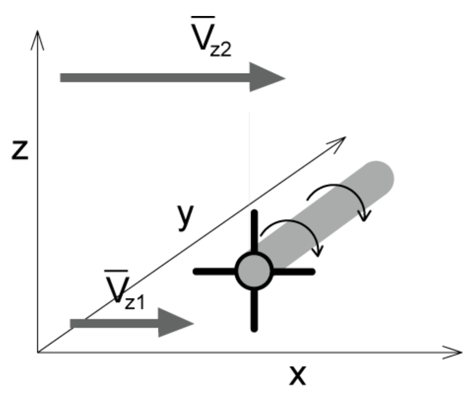
2. agestrophic circulation advects heat and kinetic energy into the frontal zone, generating vorticity and thereby further strengthening the temperature gradient

14.1.5. Frontogenesis in the upper troposphere   Martin, p. 211 – 220

Frontogenesis occurs not just in the lower and mid-troposphere, but also near the tropopause. Tropopausal frontogenesis is related to jet streams. These ‘upper-level fronts’ do not separate air masses with different origins in the horizontal, but stratosphere and troposphere.

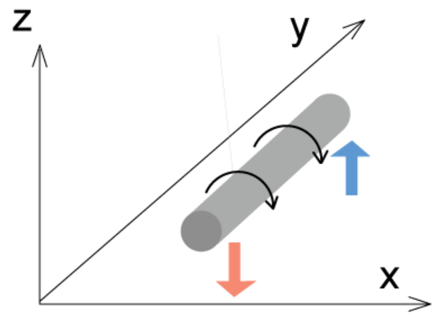
According to the thermal wind law, jet streams have air columns with horizontal temperature gradients below them. However, since wind speed increases with height below jet streams, there is wind shear below them as well, and this produces horizontal vorticity.

This can be demonstrated with a horizontal, cylindrical volume of air, which, looking along the y-axis, rotates clockwise due to the influence of shear:

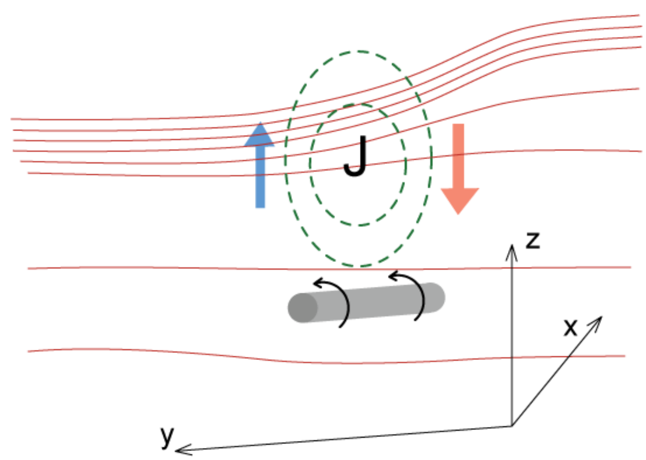


In this figure, the lower level wind V(z1) is weaker than the upper wind V(z2).

If this cylinder lies near the exit region of a jet stream, it is surrounded by thermally indirect vertical circulation (rising cold air, descending warm air):



Consider the exit region of a jet stream. The following figure has cold northern air on the left (direction y) and warm southern air on the right. In a westerly flow, the wind goes into the paper. In these conditions the cylinder contains positive vorticity:



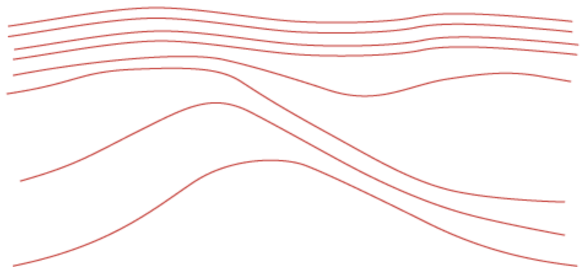
Vertical circulation brings very stable air into the troposphere from the stratosphere. θ–surfaces are close to each other in this air.

Conversely, there is a flow of less slabile air from the troposphere to the stratosphere. This causes θ– surfaces to tilt vertically, lessening the horizontal distance between them. At the same time, the cylinder itself tilts, causing the vorticity to gain vertical component.

So, there is an area around the left exit region of the core of the jet stream that contains a horizontal maximum of temperature gradient and static stability, as well as a maximum of the vorticity's vertical component. This means that there is frontogenesis. Although the exit region contains diffluence, which by itself causes frontolysis, the frontogenesis generated by the circulation is stronger.

Tropopause folding

The aforementioned circulation alters the θ field near a jet stream's core as follows:

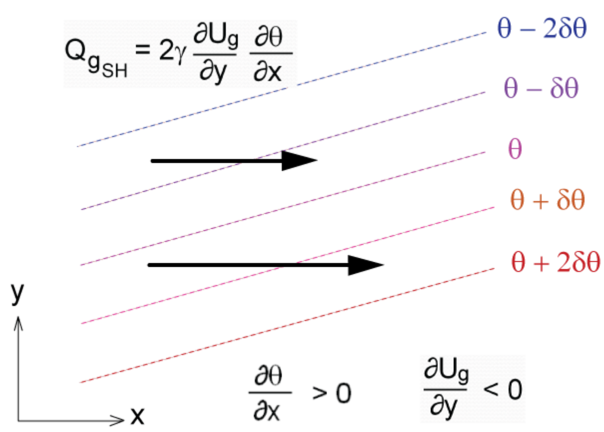


A frontal zone reaching down into the troposphere forms on the cold side of the jet stream core. The tropopause folds below the jet, and the dry intrusion is arized.

14.1.6. Frontogenesis and temperature advection in the upper troposphere

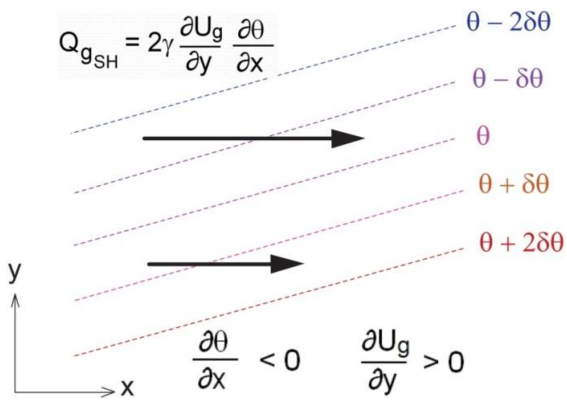
As seen in chapter 14.1.3., shear deformation generates frontogenesis if the temperature field has a favorable orientation relative to the flow. There are four different alternatives:

1. cold advection in cyclonic shear:



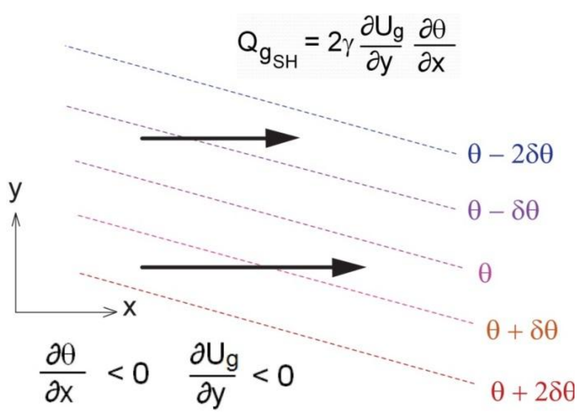
Q(GSH) < 0, thermally indirect circulation, frontogenesis

2. cold advection in anticyclonic shear:



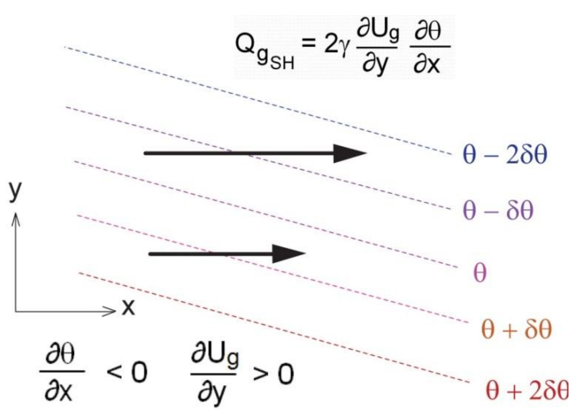
Q(GSH) > 0, thermally direct circulation, frontolysis

3. warm advection in cyclonic shear:



Q(GSH) > 0, thermally direct circulation, frontolysis

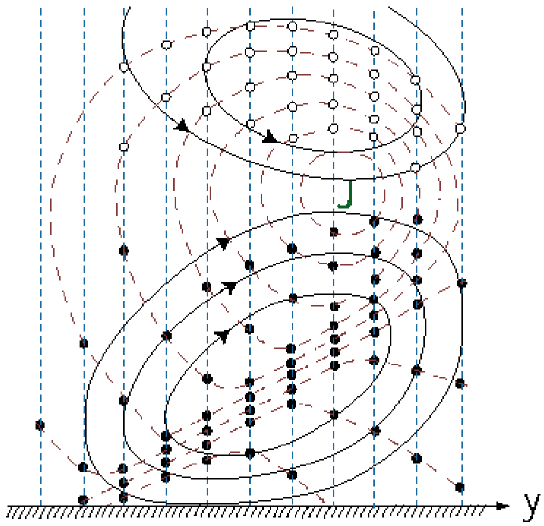
4. warm advection in anticyclonic shear:



Q(GSH) < 0, thermally indirect circulation, frontogenesis

14.1.7. Humidity and precipitation in fronts

A schematic of circulation in a frontal zone (Eliassen 1962):

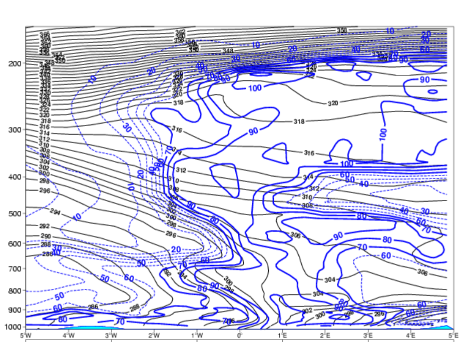


Ageostrophic circulation is marked in black, the y-component of geostrophic wind in blue dashes and the x-component of geostrophic wind in brown dashes.

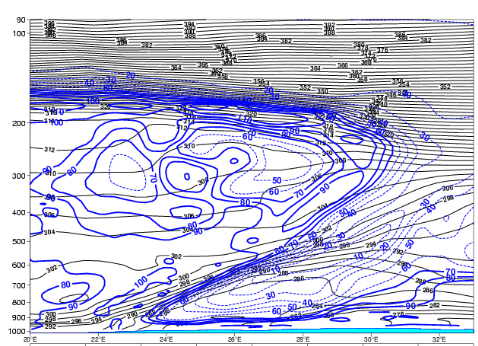
There is descending motion in the upper troposphere in both the front itself and its warm side. The front is initially dry high up, but gradually humidity increases as the ascending motion reaches higher and higher.

When polar air takes on moisture in the lower troposphere and mid-latitude air dries in the upper troposphere, clouds higher in the front no longer follow the front's inclination; the cloud layer will typically end up being roughly twice as steep as the front.

Typical relative humidity in fronts:



cold front

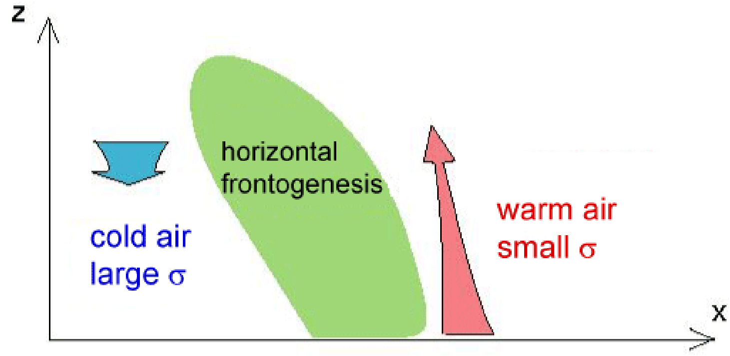


warm front

The effects of static stability:

In both the quasi-geostrophic ω-equation and the Sawyer-Eliassen circulation equation, synoptic scale ascending motion is related to static stability σ: when σ is small, the ascending motion can intensify, whereas a large σ counteracts ascending motion.

As implied by the continuity equation, mass cannot disappear, so there must be a descending flow to balance the ascending motion. A cold frontal zone:

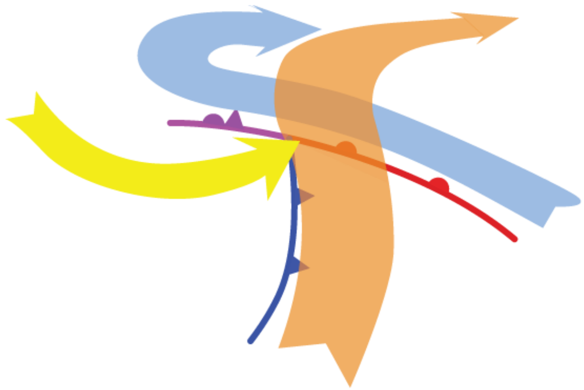


Ascending motion generated by fronts occurs in a relative narrow volume, but the descending motion corresponding to it is spread out over a large area. This is because air is fairly stable in areas with descending motion, and stable air opposes vertical motions.

Fronts undergoing frontogenesis contain rainbands, which are products of change in static stability across the front.

Convective or potential instability

Potentially unstable conditions result from warm advection in the lower troposphere and cold advection in the upper troposphere. Circumstances like this occur in developing lows near the occlusion point, where a warm conveyor belt rises and a dry, cold upper flow descends on it from the stratosphere:



The blue flow is the cool flow from the air mass preceding the cyclone, and of the three it lies closest to the surface. The orange flow is a warm flow from a warm air mass, and the yellow arrow is an upper flow.

In addition to other ascending motions, the vicinity of an occlusion point often sees the development of convection, which is why the heaviest precipitation in a cyclone are found in those areas - with the possible exception of strong showers in the cold front.

14.2. Cold fronts

14.2.1. Generally

When a cold front passes by a given location, at first the skies become cloudier in the area, which is followed by rains, a sharp change in wind direction and the wind becoming gusty. Then the rains stop and the air cools. Fronts usually travel at a rate of 20-50 km/h, but can reach speeds of up to 80 km/h.

Intense cold fronts typically have a narrowish band of uniform precipitation near the center of the low, but otherwise the rains are mostly convective.

The nature and intensity of the precipitation depends on the cyclone's stage of development, the distance from the center of the low, and the humidity of the air masses.

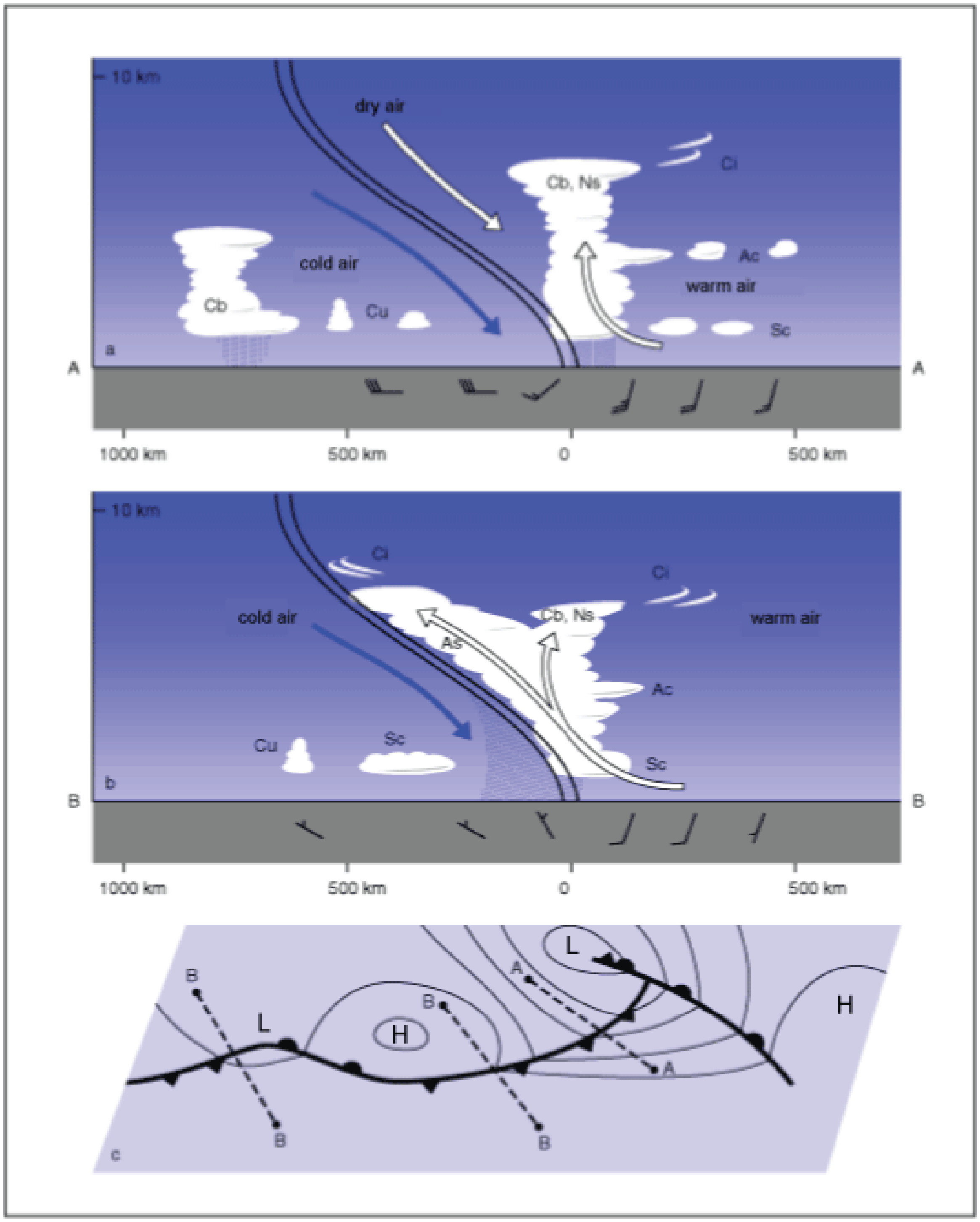
In summer, the cold front of a cyclone coming from mid-latitude air mass (usually south or southwest) can be very strong, with large temperature differences and high moisture content in the warm air mass. Heavy thundershowers may occur among or in addition to continuous rains.

In winter, if the air behind a cold front arriving from the north is very cold and dry, the front may be completely dry, even cloudless. Only changes in temperature and wind are observed.

Wintertime thunders (which are often so weak that only lightning location sensors can spot them) are related to cold fronts.

A front's intensity generally depends on its distance from the center of the low and the cyclone's stage of development. Cyclones that come to Finland from the Atlantic are usually occluded or in the early stages of occlusion.

Karttunen et al: *Ilmakehä, sää ja ilmasto*: cold front



14.2.2. Kata- and anafronts

The traditional (Bergeron 1937) way of classifying cold fronts is a division into katafronts and anafronts. In Finland they have also been called fast and slow fronts. This division is by no means always clear, and what is more, anafronts can change into katafronts. However, in those cases where the front type is easy to distinguish, it provides information on the intensity of precipitation in different parts of the front. Clear cases like this are easiest to spot in satellite images.

The differences between ana- and katafronts can be explained by the relative positions of the cold front and the warm conveyor belt:

In a katafront the warm conveyor belt travels ahead of the cold front, and frontal precipitation occur before the surface front arrives there (before the wind turns and the air cools).

In an anafront the warm conveyor belt crosses the cold front and penetrates into the territory of a cold air mass, and precipitation comes in or behind the surface front.

a.) Katafront **b.) Anafront**

Figure 6.17. Idealized schematics of front-relative orientation of the warm conveyor belt accompanying different types of cold front: (a) katafront and (b) anafront. Broad shaded arrow indicates warm conveyor belt; stippled region indicates precipitation bands (adapted from Browning 1990).

Katafronts

In a katafront dry and cold air flow meets the clouds produced by the warm conveyor belt, driving them before itself and drying them where it comes into contact with their upper parts. Precipitation occurs ahead of the front and is not particularly heavy.

Katafronts are also called split cold fronts because they have two regions of cloud layers with different thickness. The border between the two cloud masses is sometimes called the upper cold front. Rainfall is heaviest in the warm sector ahead of the kata front.

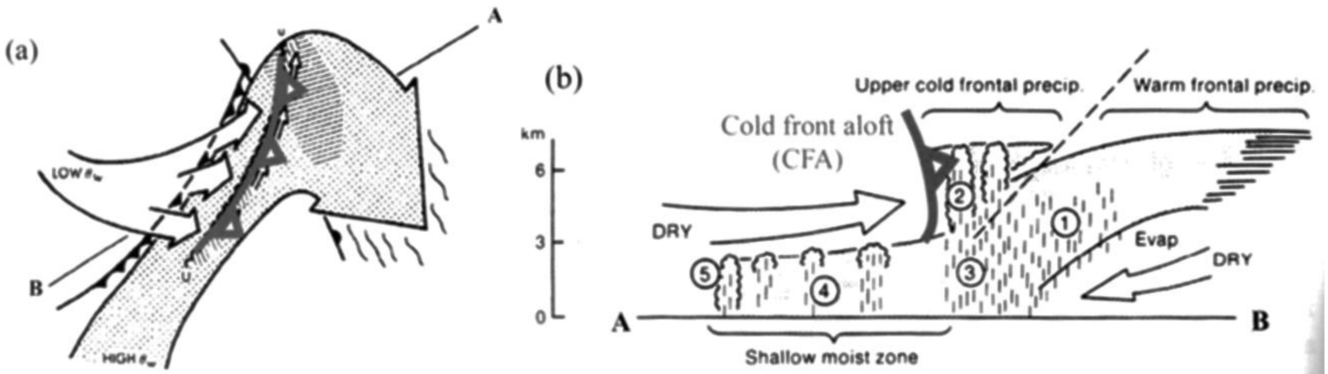


Figure 6.18. Schematic of a katafront or split-front structure with a cold-front aloft. (a) Plan view showing orientation of warm conveyor belt as large, shaded arrow, with surface fronts and indication of cool, dry low-θ_e air aloft given by hollow arrows. Line A–B indicates orientation of vertical cross section shown in (b). In (b), the numbers correspond to (1) warm-frontal precipitation, (2) convective cells ahead of the cold front aloft, (3) precipitation from upper cold front falling into lower warm-advection region, (4) shallow moist zone, and (5) light precipitation with surface cold front (from Browning and Monk 1982, their Fig. 5; Browning 1990, his Fig. 8.10).

Anafront

Dry and cold upper flow does not cross cold anafronts. Instead, the flow turns until it is parallel to the front, weakening in the process:

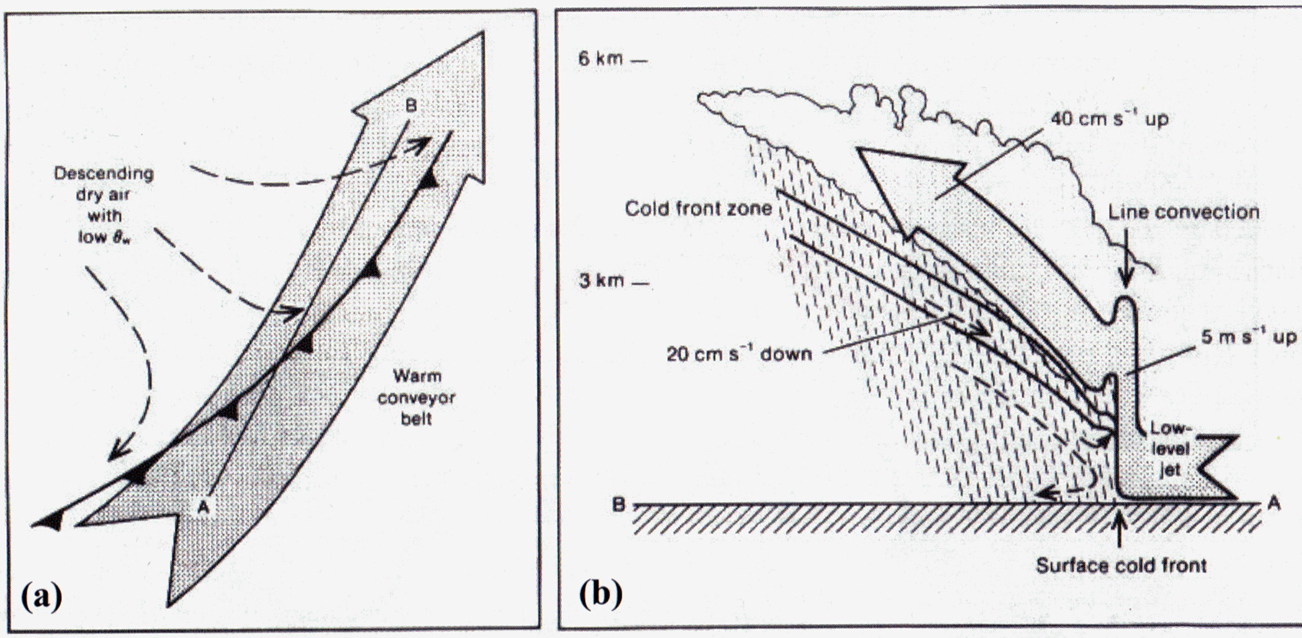


Figure 6.20. Schematic of anafront structure. (a) Plan view showing orientation of warm conveyor belt as large, shaded arrow, with surface cold front and indication of cool, dry low-θ_e air aloft given by dashed arrows. Line A–B indicates orientation of vertical cross section shown in (b) (from Browning 1990, his Fig. 8.8).

Strong convection (showers and thunder) occurs at the leading edge of the front, and there may also be a jet stream in the lower troposphere. Higher up one finds weaker ascending motion and more uniform precipitation.

Air flows and hydrometeors associated with anafronts according to Matejka (1980):

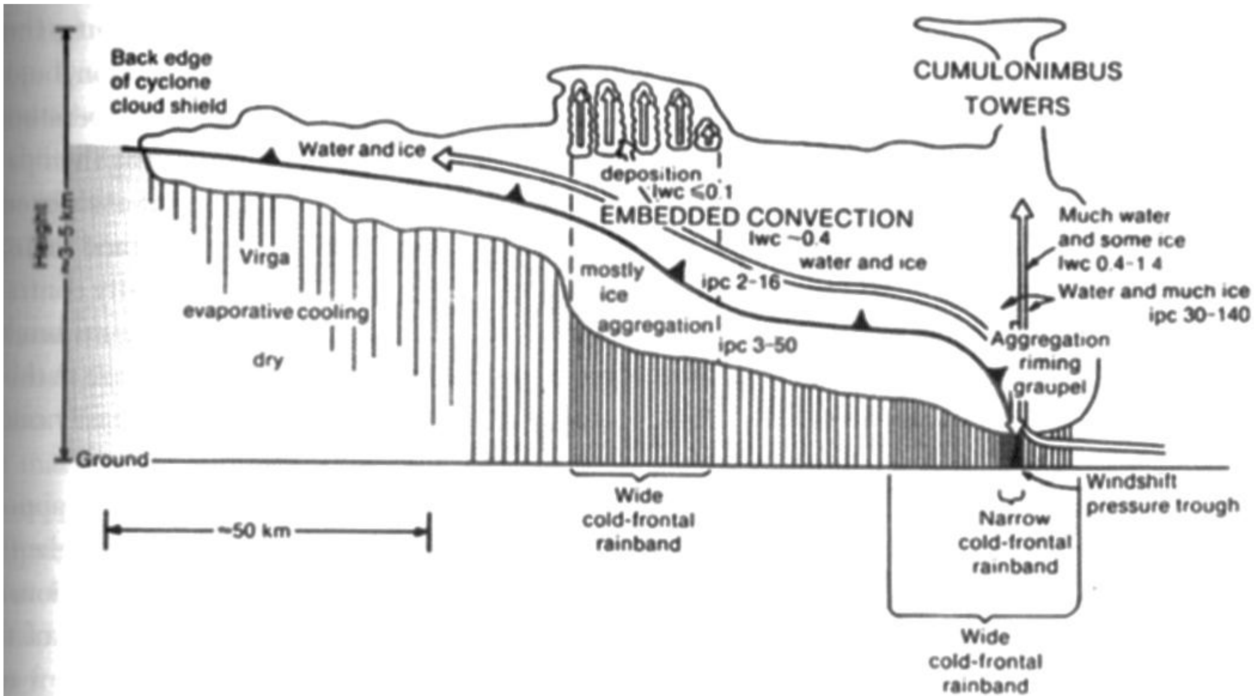


Figure 6.21. Idealized cross section of clouds and precipitation in an anafront. Vertical lines represent precipitation, and arrows represent airflow relative to front. Ice particle concentration (IPC) in pounds per liter, and liquid water content (LWC) is given in g m^{-3} at select locations. Figure taken from Browning (1990), originally from Matejka et al. (1980).

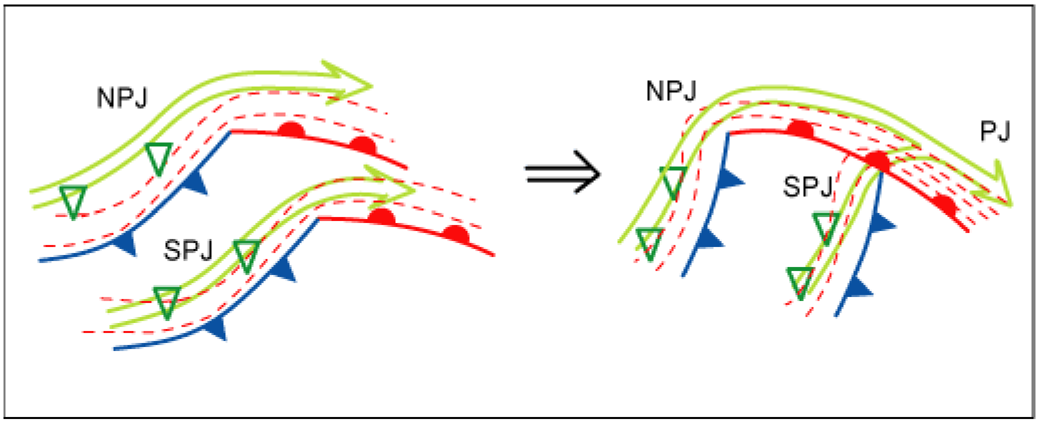
Cold fronts are typically ana type in powerful cyclones still in development. An anafront often transforms into a katafront when a cyclone occludes, as the cold, dry flow strenghtens and approaches the cold front. Both types can exist at the same time in different parts of the front.

Backdoor cold fronts:

Backdoor cold fronts come from an unusual direction, in Finland from east-northeast. This can happen if a cyclone circles clockwise around a high pressure system. When this happens, the cyclone brings in very cold air from the Arctic Sea, causing a dramatic drop in temperature.

Sometimes a cyclone may have two cold fronts:

Two polar jet stream branches, each with its own cyclone, meld together. The warm front of the southern system catches up to the warm front of the northern one. There is widespread cold advection, and two separate zones of temperature gradient can be distinguished:



14.3. Warm fronts

A warm front bring with it increased cloudiness followed by rain, and finally a change in wind direction when the rains cease. Winds pick up temporarily when the front approaches, but unlike with a cold front, it does not become gusty. The winds are strongest slightly ahead of the front, and they turn slowly over a period of about half an hour to two hours. If the front is extensive and the precipitation last a long time, frontal fog may develop ahead of it. In winter there is often freezing drizzle or rain in the warm front.

Turbulent mixing causes warm fronts to move more slowly than cold fronts. Turbulence develops when wind shear is high enough and static stability sufficiently low.

Troughs in warm fronts usually do not reach the depth they do in cold fronts, and changes in wind direction are not as sharp. Sometimes the wind direction changes associated with warm fronts - particularly ones that are old or far from the center of the low - are so weak that they are difficult to analyze on surface charts.

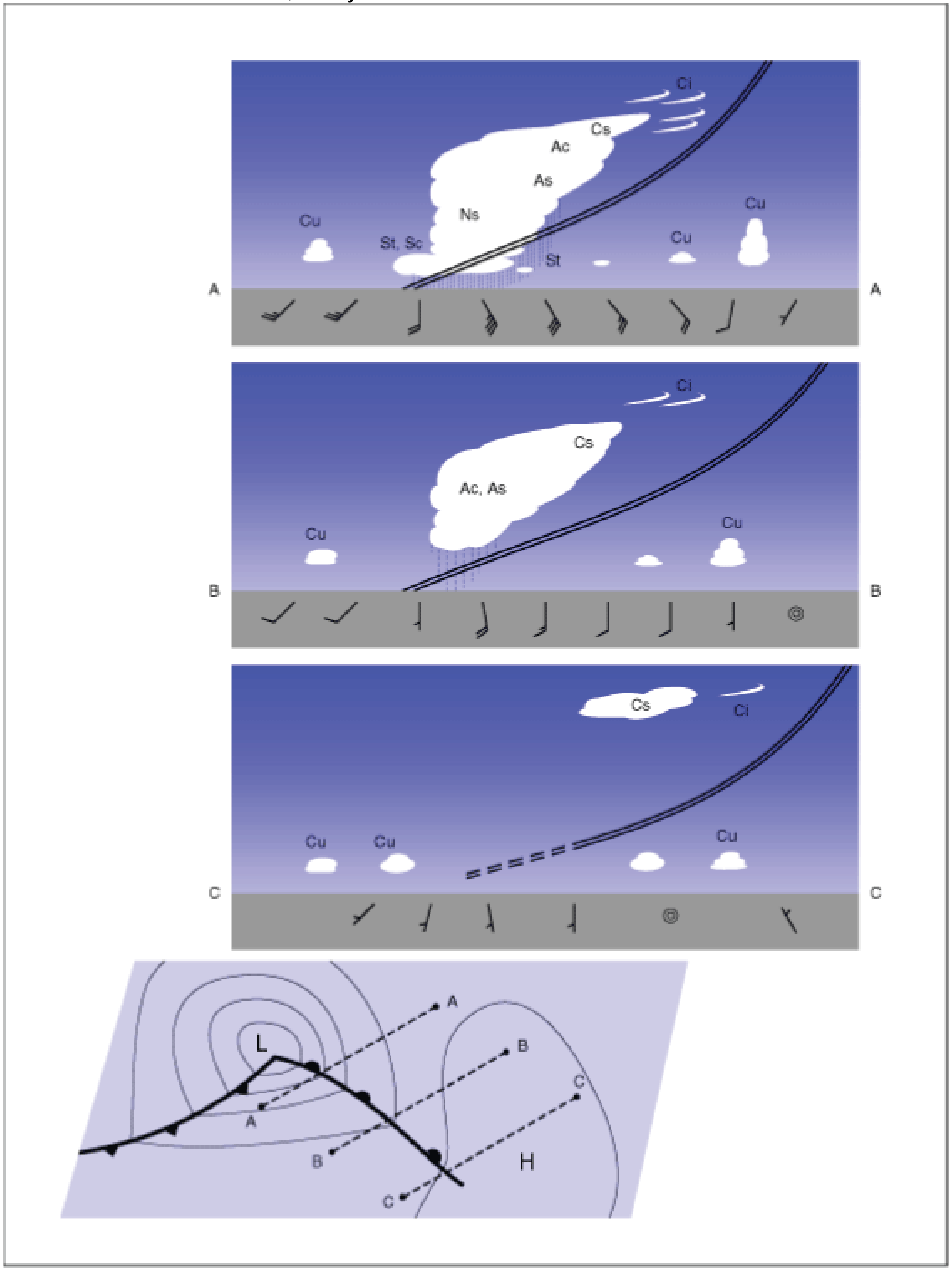
Backdoor warm fronts

Backdoor warm fronts may arrive in Finland from the east in summer. The winds preceding those fronts are northerly, bringing cold air from the Arctic Sea. On the other hand, air in the sector may be very warm; the temperature can rise as much as 20 °C after the front has passed.

Masked warm fronts

In winter, the temperature may drop after a warm front has passed. If the skies are clear before the front comes, there will be a surface inversion with low temperatures near the ground. When the front arrives, its winds destroy the inversion and the temperature rises. If there is fog, stratus clouds and/or wind in the warm sector, the temperature will remain higher than it was before the arrival of the front.

Karttunen et al: *Ilmakehä, sää ja ilmasto*: warm front



Warm fronts can be classified according to the type of their temperature gradient and appearance in satellite images. There are three types: band, shield and detached warm fronts.

Warm front band

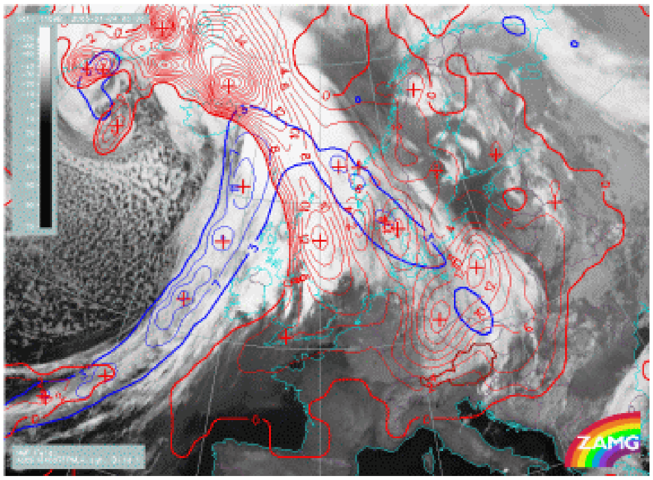
- tight temperature gradient and fairly narrow areas of clouds and rain
- relatively well-defined front
- no precipitation behind the front
- can change into the shield type if the cyclogenesis is strong

Warm front shield

- temperature gradient is widespread
- the location of the surface front may be difficult to pinpoint
- changes in wind direction not as clear as with the band type
- widespread precipitation, often also in the warm sector
- the warm conveyor belt is stronger than in the band type cyclone
- sometimes there are two warm advection maximums: the front is 'double-barrelled'

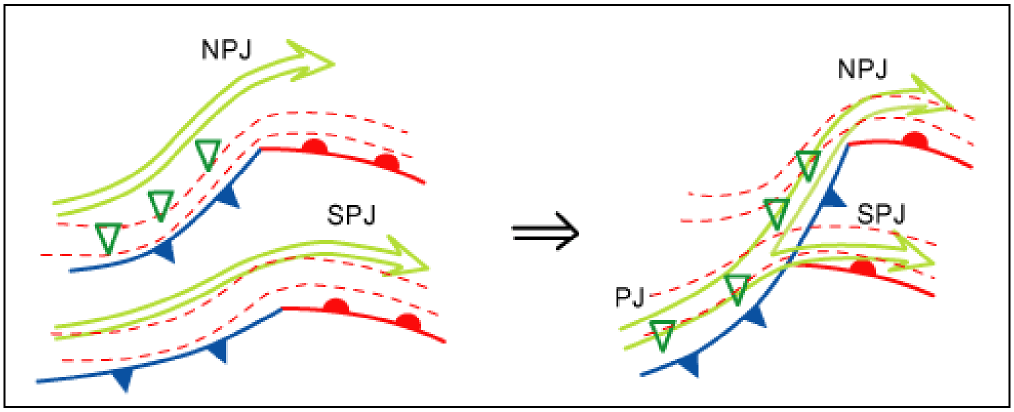
Detached warm front

- the front is often band-type, with a uniform temperature gradient
- appears to have been cut in two at the middle; the front is invisible on the surface, the cloud cover is thinner and there is no precipitation
- forms to the east of a strong high pressure system or ridge as a cyclone circles around the high from the north



Two warm fronts:

Two warm fronts may develop in one cyclone if two jet stream branches, each with their own cyclone, merge. A polar jet stream branches into two, and the northern cyclones's cold front reaches that of the southern one. Although the warm advection is spread over a wide area, two temperature gradient zones can be distinguished:



14.4. Occluded fronts

14.4.1. Warm occlusion

The word occlusion comes from the Latin 'occludere', which means 'to close up'. It refers to the closure of the warm sector. A cyclone transfers warm air polewards from the equator; in Finland this means that mid-latitude air masses southern polar air masses move north. In the occlusion stage, all the warm air has moved away and only cold air remains around the occluded front.

The Norwegian model divides occlusions into warm, cold and neutral occlusions. In practice it is difficult to distinguish the type of most occlusions. Because the surface has a large effect on 2 m temperatures, an occlusion's type is primarily determined by 850 hPa temperature (or 750 hPa in mountainous regions).

Warm occlusion: the air is warmer behind the occluded front than ahead of it.

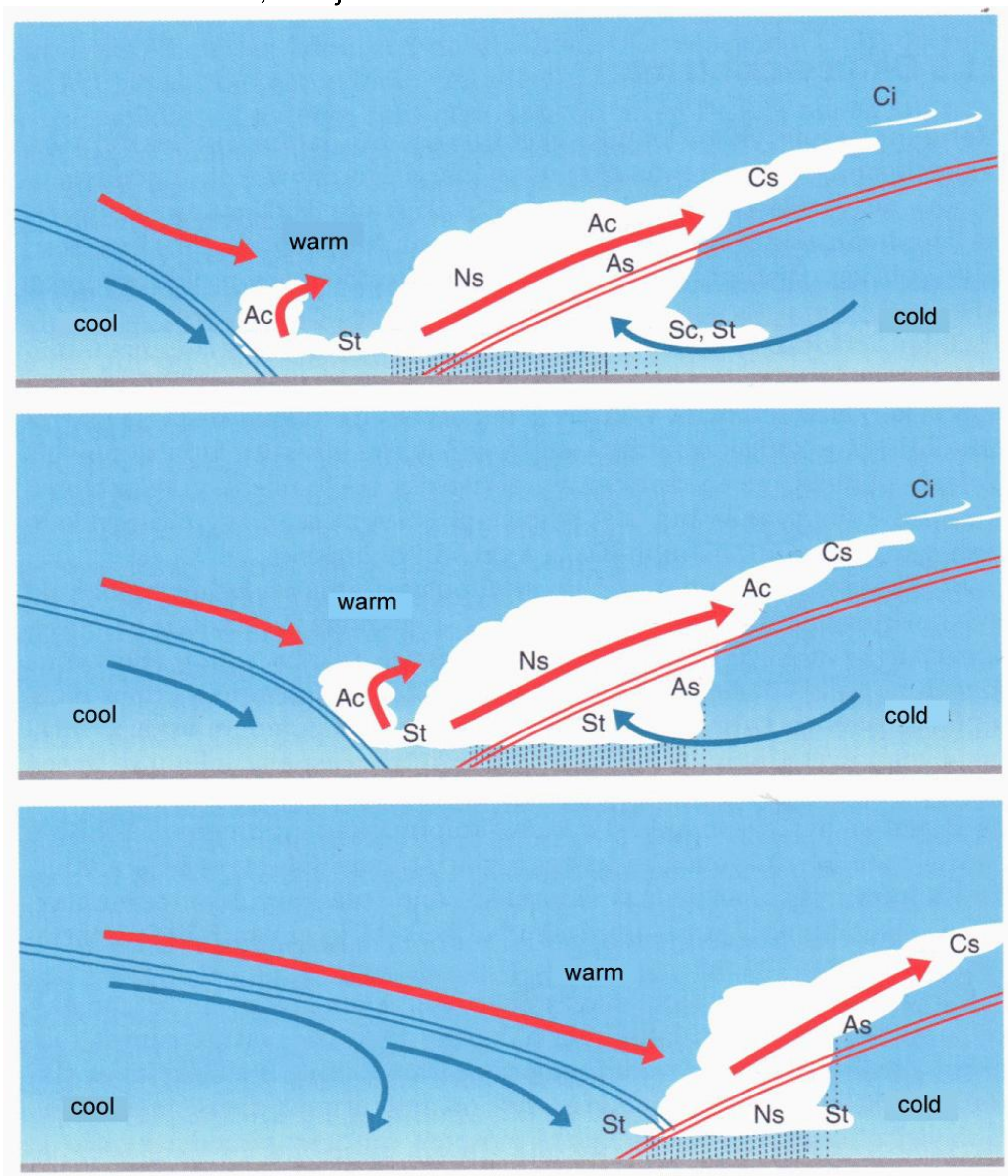
In winter, occlusions that come from the Atlantic are typically warm due to marine air masses being warmer than continental ones. These occlusion types are the clearest.

A warm occluded front resembles a warm front, but it contains more convective cloudiness and is scattered in its later stages.

In Finland the heaviest snowsfall comes from strong warm occluded fronts that are preceded by southeasterly winds and heavy precipitation. Temperatures may rise up from -20°C to around 0°C after the front has passed. Warm occluded fronts, like warm fronts, can produce freezing rain or drizzle.

In summer the temperature difference between the two sides of a warm occluded front is usually small.

Karttunen et al: *Ilmakehä, sää ja ilmasto*: warm occlusion:

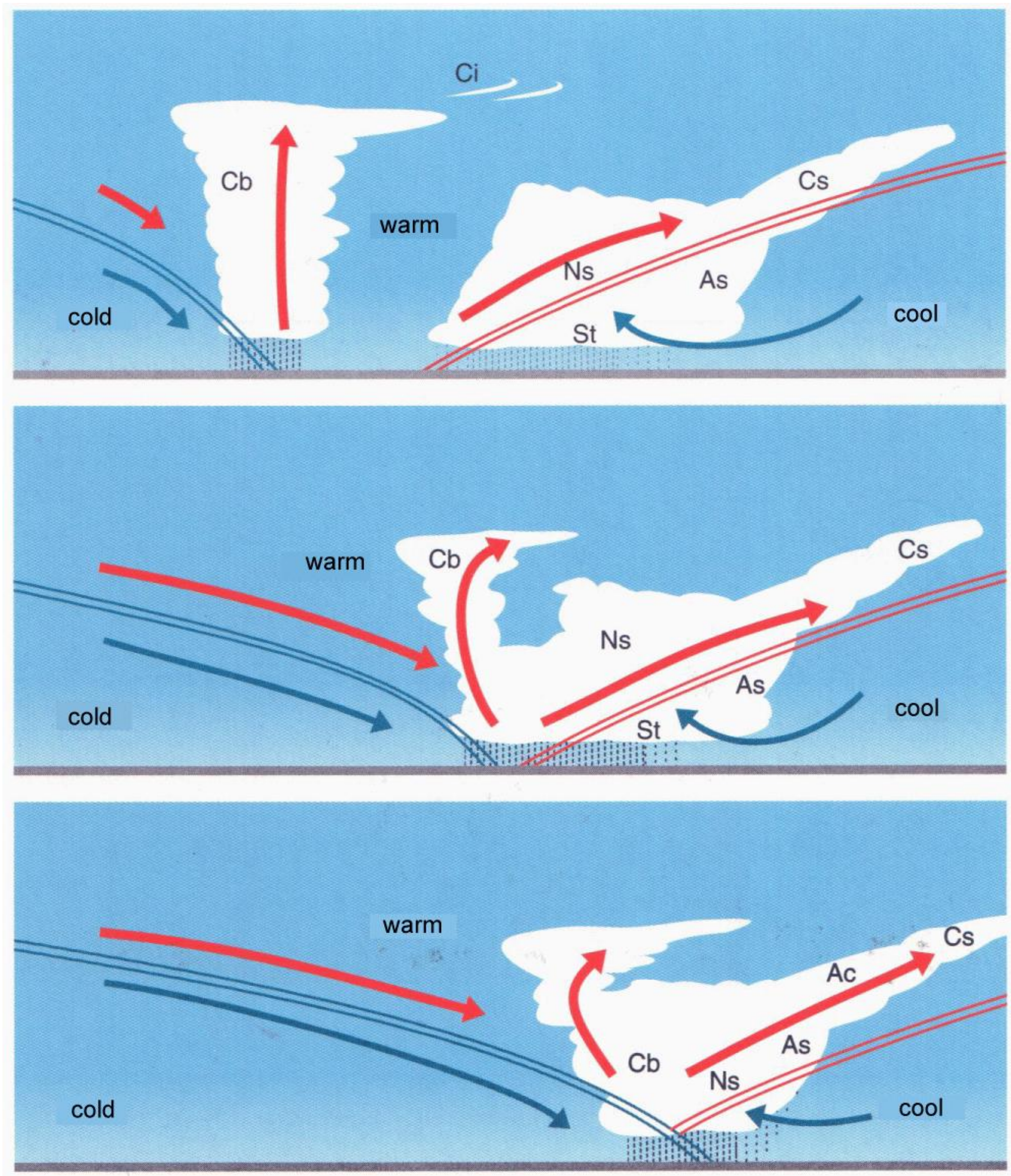


14.4.2. Cold occlusion

Cold occlusion: the air is cooler behind the front than ahead of it.

In Finland cold occlusions are most common in summer, when the sea is colder than the continent. Cold occlusions can resemble cold fronts in extreme cases.

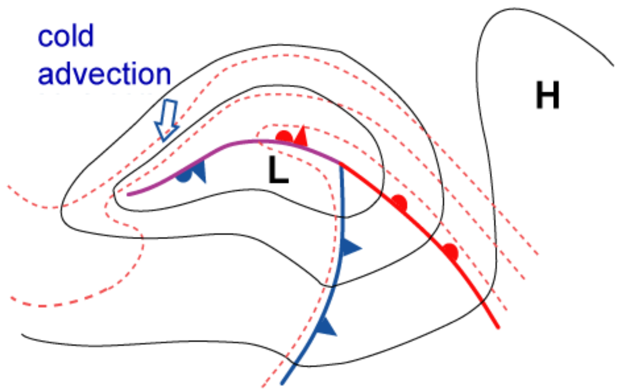
Karttunen et al: *Ilmakehä, sää ja ilmasto*: cold occlusion:

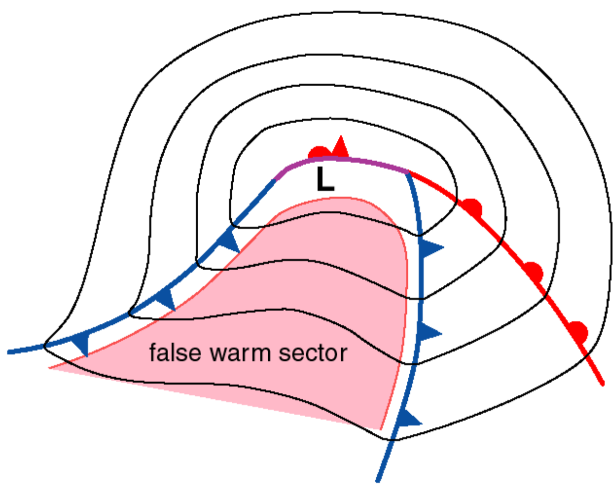


With neutral occlusions, the air has the same temperature on either side of the front. Such an occluded front forms a vertically straight wall at the surface. The lower reaches of most cold occlusions are also nearly vertical because friction slows the front down near the surface.

14.4.3. Bent-back occlusion

If there is a flow of cold air behind an occluded front and it originates from the eastern parts of an upper high located to the north of the front, the upper high's tail might move off with the flow in the opposite direction from the rest of the front. Such bent-back part of an occluded front often transforms into a cold front and may surpass the intensity of the original cold front.

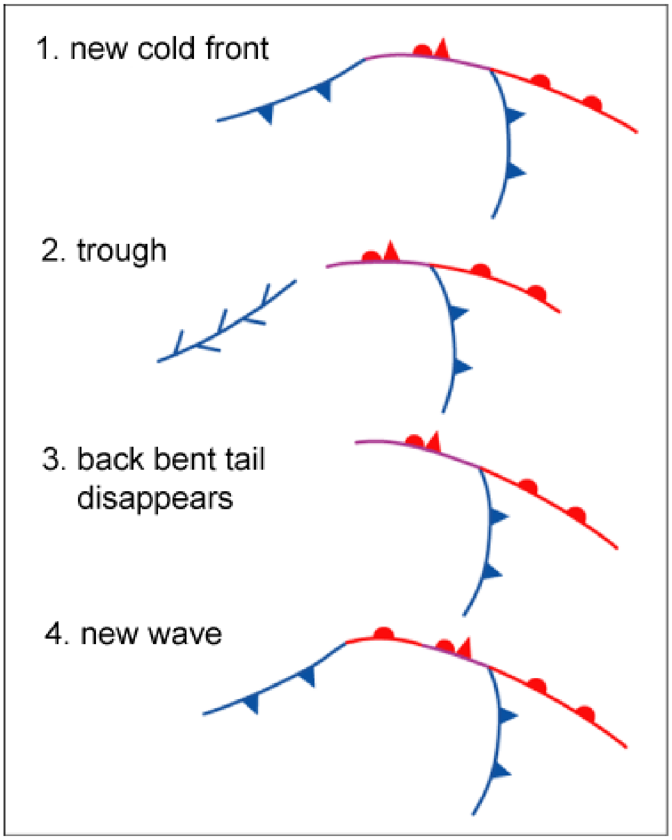




Bent-back occluded fronts are worth keeping an eye on, as they can have significant effects on the weather:

- drop in temperature
- showers
- gusty winds

The four possibilities for further development:



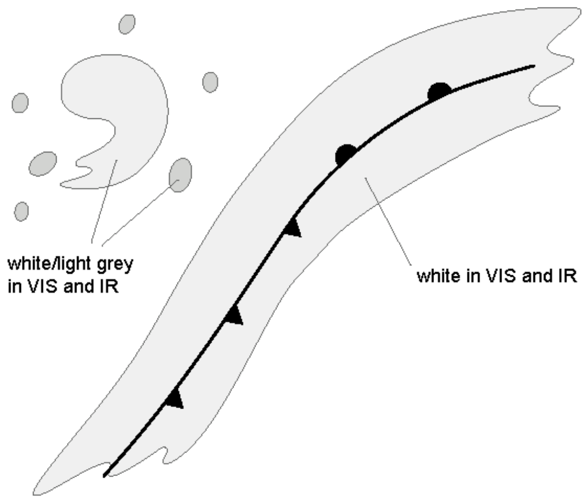
14.4.4. Instant occlusion

Instant occlusion is a cloud system that gets it name from a temporary resemblance to an occlusion in satellite images. Although they are not actually occlusions, they may develop into such over time.

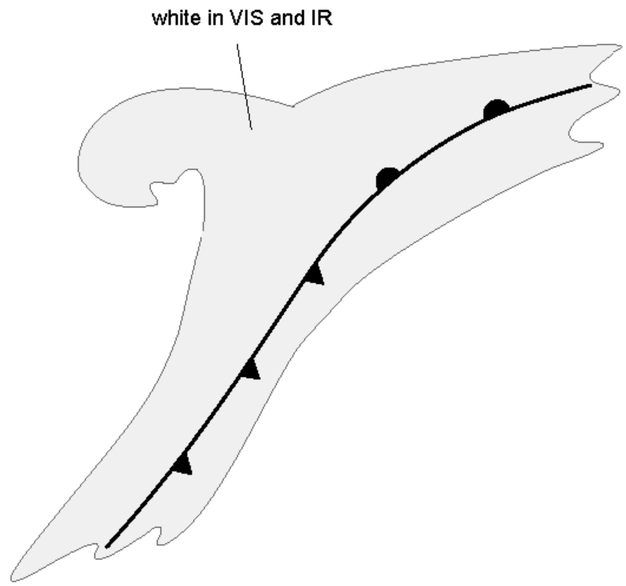
Instant occlusions occur when there is a small trough traveling along the edge of a large upper low, and a small Comma Cloud accompanying it. If a trough like this overtakes a baroclinic wave, the system may look like an occlusion. If there is a wave in its developmental stage, the upper trough will accelerate the development.

Schematics of an instant occlusion's clouds and fronts:

The starting point:



Instant occlusion stage:



14.5. Arctic fronts

Arctic fronts form at the border of arctic and northern polar air masses. They can reach Finland in wintertime, but they rarely make it further south than the 60th latitude.

Warm and occluded fronts are so weak in arctic cyclones that they are practically invisible. As a result, an arctic cyclone is often simply called an arctic cold front. The conveyor belts associated with arctic cyclones are likewise much less clear than ones in polar air.

A low-lying baroclinic zone that develops where frozen and ice-free sea meet can also be called an arctic front. It may head south with the basic flow and a great deal of low convective clouds develop in its wake.

Arctic fronts are shallow; they can only reach heights of 850-700 hPa. They are accompanied with an arctic jet streams, but those are often too weak to detect on pressure charts, though they are still visible in cross sections. There is arctic air behind an arctic front, and polar air above the arctic air. Winds turn sharply with the passing of an arctic front.

The temperature behind arctic fronts is -18 ºC or less. If, following the passing of the front, the weather clears and a strong high pressure system or ridge develops (allowing the surface to cool and a storng surface inversion to develop), the temperature reaches very low values. In that case the front is often dry or even cloudless, and there is little to mark its passing except the drop in temperature.

Humid marine arctic fronts bring snow showers with them, followed by low convection after the showers have passed.