Dynamic Meteorology: lecture 11

21/12/2022 (Wednesday, 13:15)

Sections 1.31 (part G) and 1.37-1.39 (part H)

Quasi-balance and quasi-geostrophic approximation (section 1.31)
Rossby waves (section 1.37)
Excitation of Rossby waves due to baroclinic instability (section 1.38)
Baroclinic life cycle: Norwegian cyclone model (section 1.39)

Part F is skipped *. This part is meant as background reading and inspiration for **project 5.**
Sections 1.32 and 1.33 (on tropical cyclones) are postponed to after Christmas.

Tutorial after this lecture (11)
Project 5
Problems 1.20, 1.21 (part E) p. 153
Problem 1.22 (part F) p. 189
Problems 1.34 and 1.35 (part G) p. 252-253

* except problem 1.22

(a.j.vandelden@uu.nl)
(http://www.uu.nl/~nvdelden/dynmeteorology.htm)

Quasi-balance in the atmosphere

This lecture is an introduction to exercises 1.34 and 1.35 on Rossby waves !

Quasi-balance and quasi-geostrophic approximation (section 1.31)

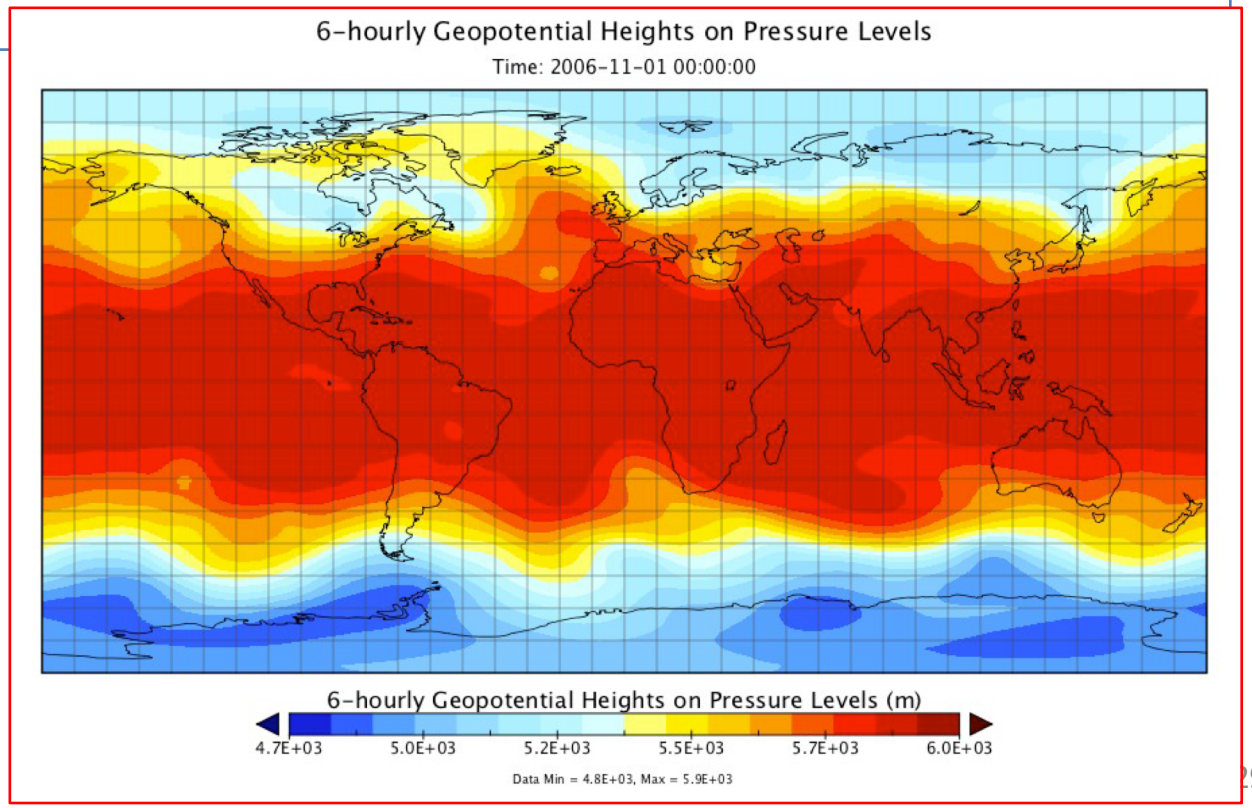
Rossby waves (section 1.37)

Excitation of Rossby waves due to baroclinic instability (section 1.38)

Baroclinic life cycle: Norwegian cyclone model (section 1.39)

Next lecture (12): Balanced vortex (tropical cyclone) and potential vorticity inversion (sections 1.32 & 1.33)

Rossby waves in the geopotential at 500 hPa
November 2006-February 2007



ERA-Interim re-analysis
h500Winter2006-2007.mov, made with Panoply

Remaining lectures and topics in 2023

Lecture 11: 21/12: quasi-balance; quasi-geostrophic approximation (part G), planetary (Rossby) waves, Norwegian cyclone model, baroclinic instability of Rossby waves and cyclogenesis (part H)

Lecture 12: 11/1: balanced vortex (tropical cyclone), potential vorticity inversion applied to an axisymmetric vortex (part G)

~~**Lecture 13:** 13/1: poleward mass flux due to isentropic potential vorticity mixing by breaking Rossby waves (part H)~~

Lecture 13: 13/1: Fronts, frontogenesis, Q-vector, omega equation (part I)

Lecture 14: 18/1: Omega equation: vertical motion in Rossby waves

20/1: Practicing multiple choice questions

Oral presentations of project 5: Wed. 18/1/23 and 25/1/23

Exam: Wed. 1 Feb. 2022, 13:30-16:30 - EDUC-BETA

Retake: somewhere in April

Summer convection: **severe imbalance**

The wind direction has no relation with the isobars! This situation is in fact quite exceptional !

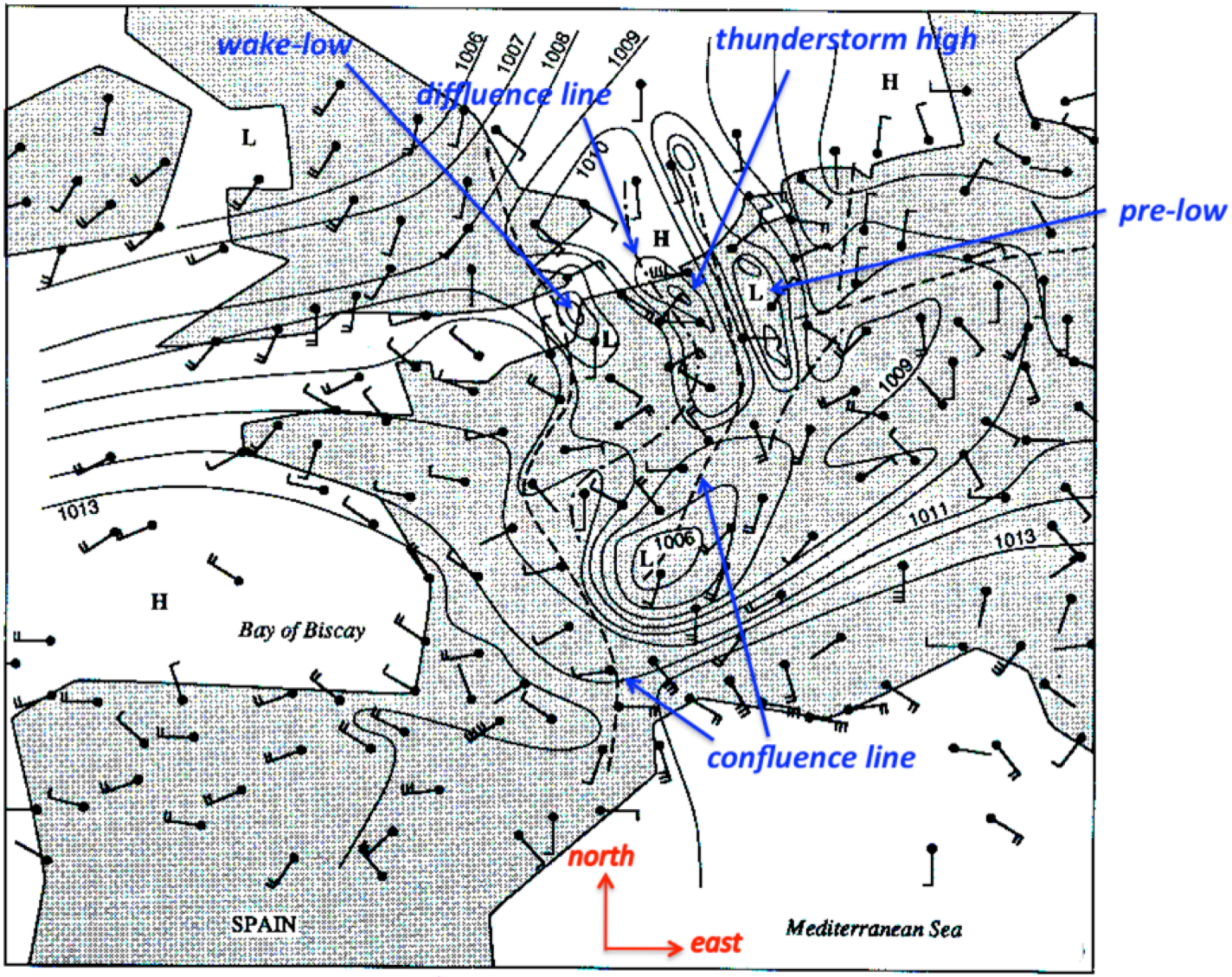
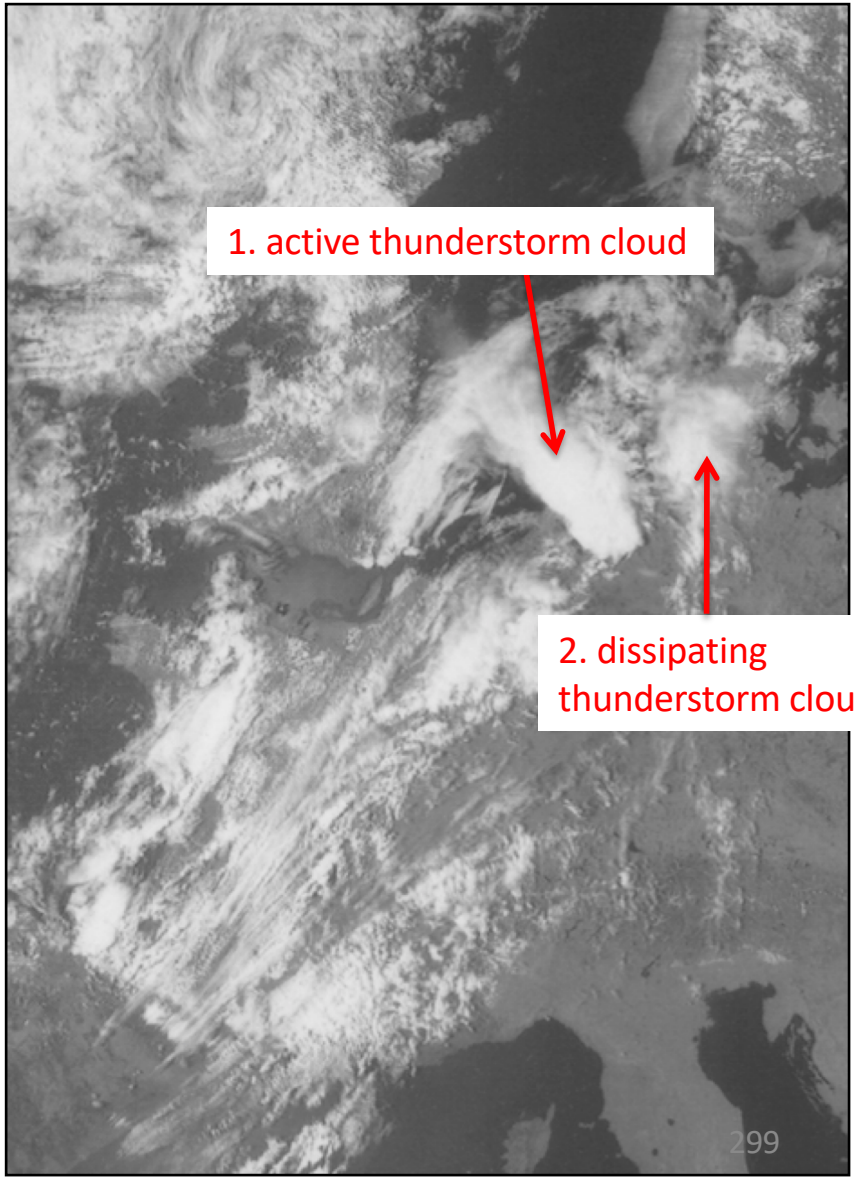


Figure 1.90: Sea level pressure distribution and surface wind associated with a thunderstorm "system"

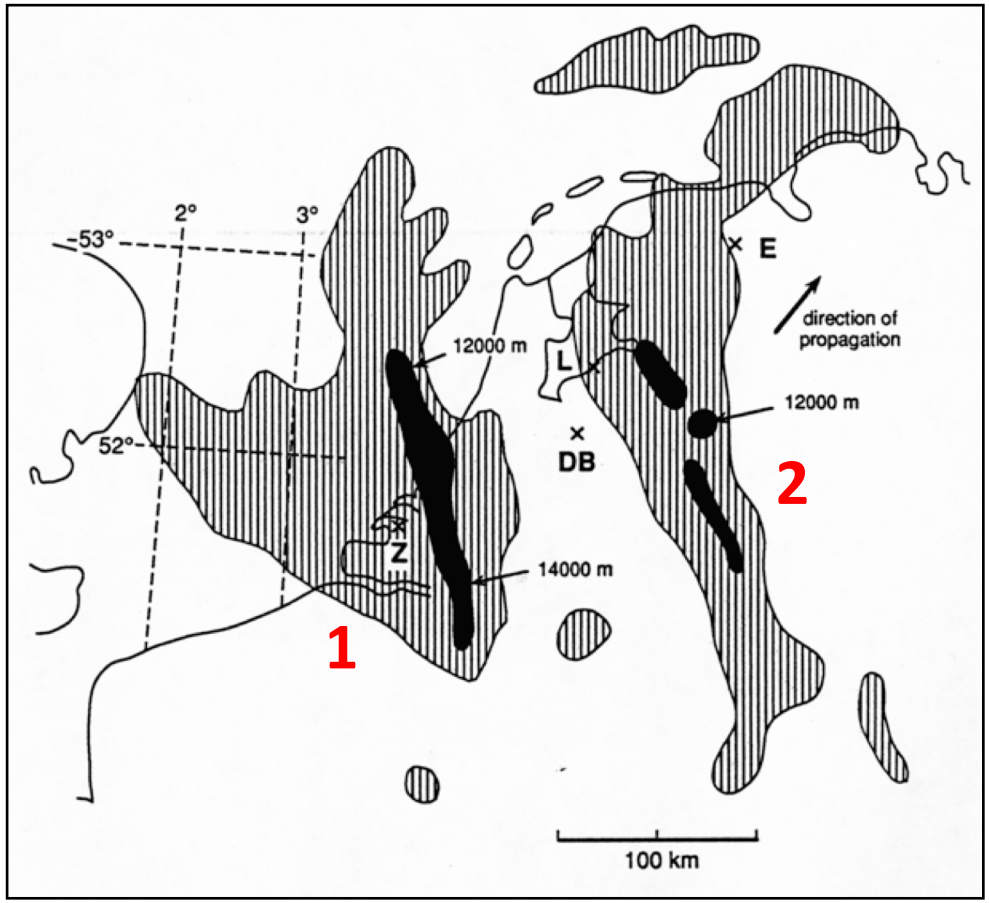
Some details of this case of severe convection

The thunderstorms labeled “1” and ”2” formed over the northern Spain and south-west France in the evening of July 10, having a life time of at least 12 hours!

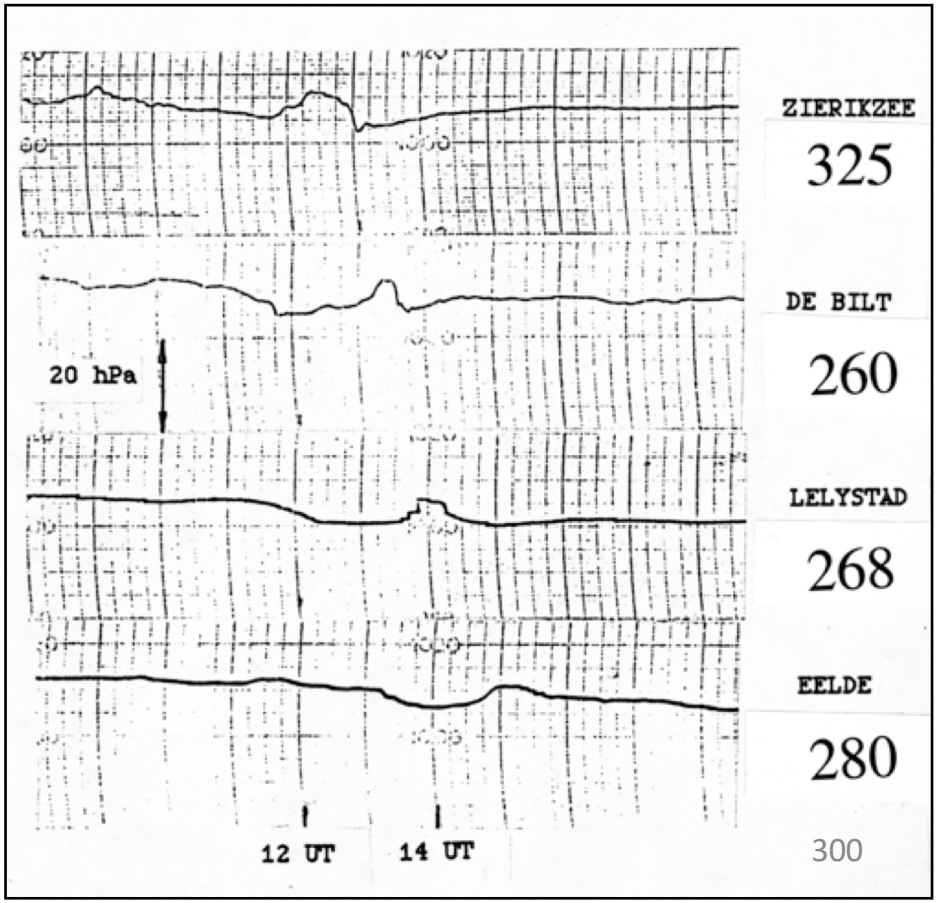
Image (VIS) of the thunderstorm clouds over Belgium and the Netherlands made by a polar orbiting satellite on July 11, 1984 at 14:29 UTC.



Schematic representation of the precipitation features in the thunderstorms derived from the radar equipment based at De Bilt (DB) on July 11, 1984, 12:32 UTC (**black: relatively intense precipitation**). The other stations indicated are Zierikzee (Z) (325), Lelystad (L) (268) and Eelde (E) (280). The complex of thunderstorms is moving in northeasterly direction. Also indicated are the maximum heights in metres of the echo's at several points.



Surface pressure as a function of time at four stations in the Netherlands along the path of the thunderstorms. See **the figure on the left** for the location of the stations.



Buys Ballot's Law



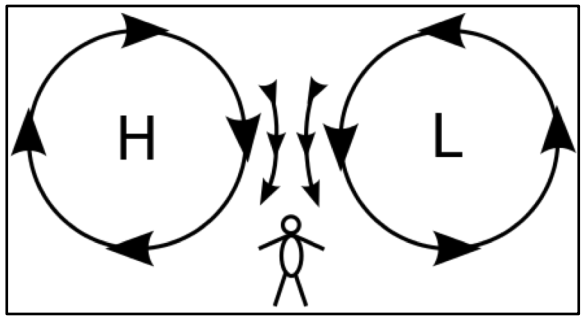
Christophorus Buys Ballot (1817-1890)

"If a person stands with his/her back to the wind in the Northern Hemisphere, atmospheric pressure is low to the left and high to the right"

If such **chaos** (**slide 298**) were the rule, instead of the exception, the Dutch meteorologist, Christophorus Buys Ballot, founder of KNMI, would not have identified a robust empirical relation between pressure gradient and wind direction, now referred to as "**Buys Ballot's law**" or "**geostrophic balance**", at the earth's surface in 1857. In *pressure coordinates* "geostrophic balance" is written as…

$$u_g = -\frac{1}{f}\left(\frac{\partial \Phi}{\partial y}\right)_p \qquad v_g = \frac{1}{f}\left(\frac{\partial \Phi}{\partial x}\right)_p$$

The *geopotential* is defined as $\Phi \equiv g z_p$



the stick figure is facing out of the picture

How close to geostrophic balance is the atmosphere?
How did Buys Ballot find his law?

Rediscovering Buys Ballot's Law

Rediscover Buys Ballot's law from observations of the same type that were used by Buys Ballot himself in the nineteenth century. We use observations of sea level pressure and wind taken at IJmuiden, Valkenburg and Schiphol, separated at distances of 15 to 35 km, at or close to the west coast of the Netherlands (**figure 1.35**),

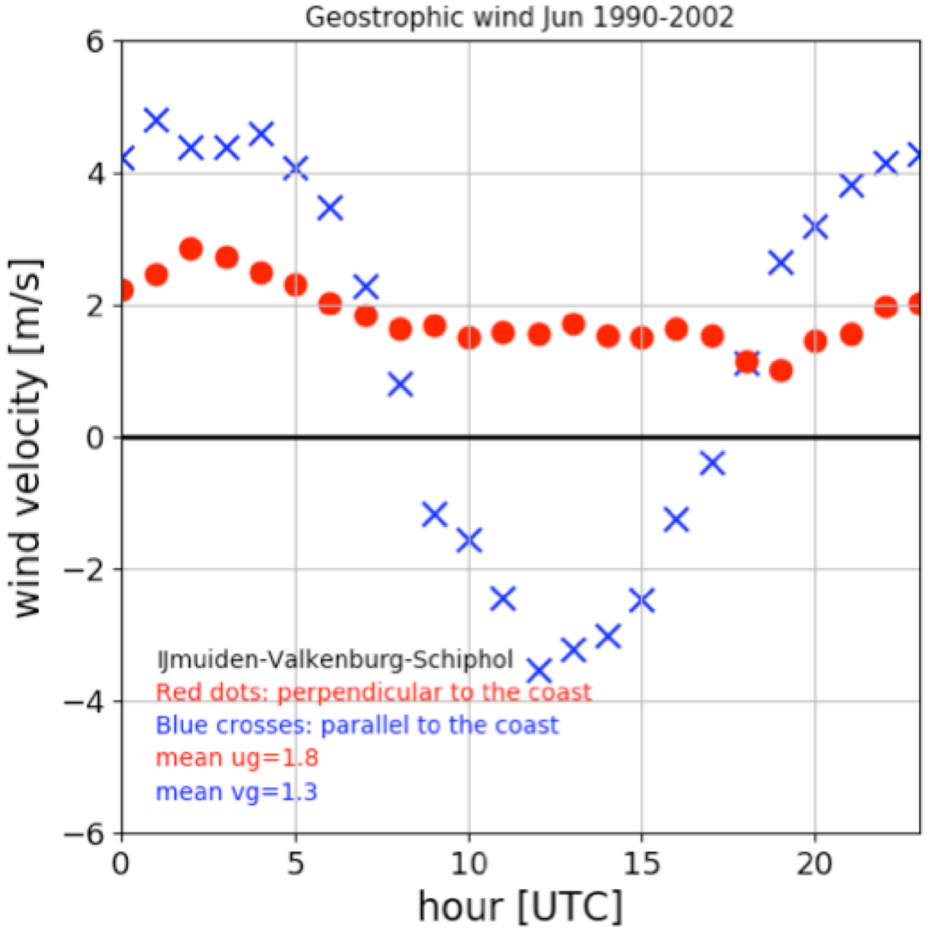


FIGURE 1.91. June (1990-2002) average daily cycle of the geostrophic wind velocity **perpendicular** (**red dots, u_g**) and **parallel** (**blue crosses, v_g**) to the Dutch west coast. Positive values of the geostrophic wind parallel to the coast, during the night, are associated with *high pressure over land* ($\partial p/\partial x>0$). Negative values, during the day, are associated with low pressure over land ($\partial p/\partial x<0$). In June sea-level pressure is relatively low over land between 9:00 and 17:00 UTC. The daily mean geostrophic absolute wind velocity of 2.2 m/s, from the south-west.

The geostrophic wind is calculated from hourly observations of sea-level pressure at Valkenburg, Schiphol and IJmuiden by the method described in **problem 1.13**.

Rediscovering Buys Ballot's Law

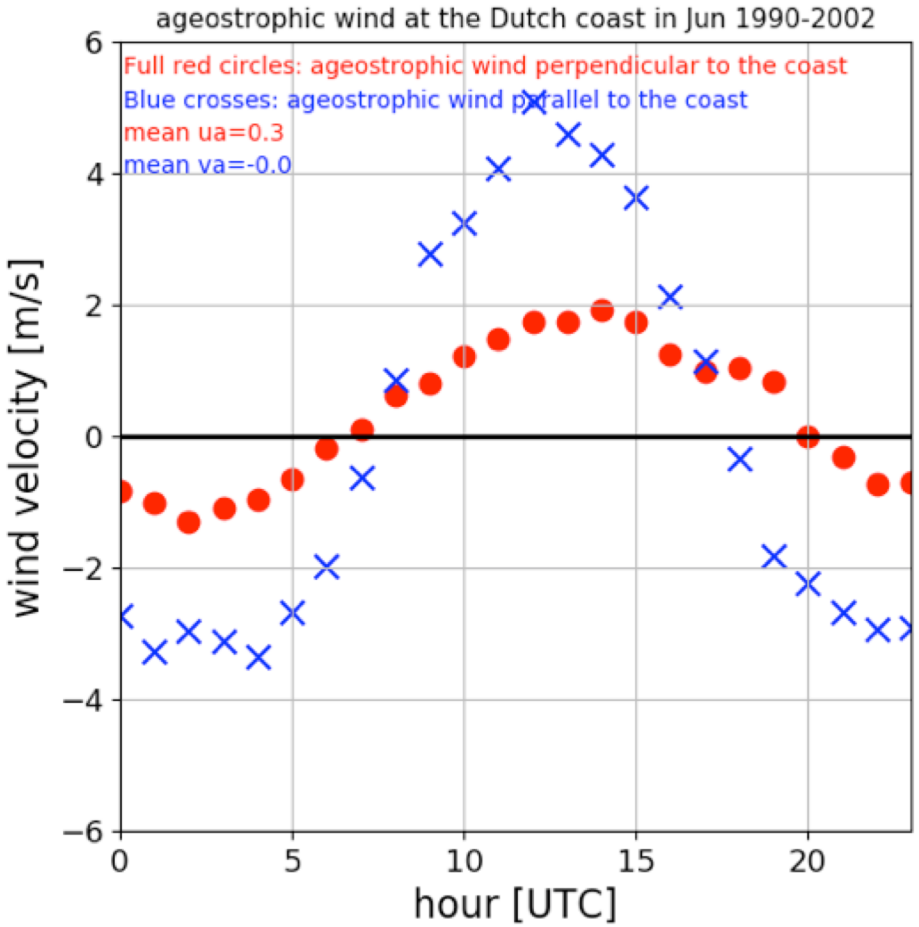


FIGURE 1.91. June (1990-2002) average daily cycle of the ageostrophic wind velocity **perpendicular** (**red dots, u_a**) and **parallel** (**blue crosses, v_a**) to the Dutch west coast. Positive (negative) values of the ageostrophic wind velocity perpendicular to the coast (**red dots**) correspond to a sea breeze (land breeze).

The daily mean absolute ageostrophic wind velocity of 0.3 m/s, which is much smaller than the daily mean geostrophic absolute wind velocity of 2.2 m/s.

The atmosphere is close to geostrophic balance only for averaging times longer than 1 day.

The daily mean geostrophic wind is 2.2 m/s
The daily mean a-geostrophic wind is 0.3 m/s

→ $|\vec{v}_a| << |\vec{v}_g|$

AnalysisSeaBreezeObservations.py

Planetary wave & cyclonic circulation: in **quasi-balance:** $|\vec{v}_a| << |\vec{v}_g|$

Buy's Ballot's law in pressure coordinates:

$$u_g = -\frac{1}{f}\left(\frac{\partial \Phi}{\partial y}\right)_p \qquad v_g = \frac{1}{f}\left(\frac{\partial \Phi}{\partial x}\right)_p$$

$$\vec{v} = \vec{v}_g + \vec{v}_a$$

geostrophic

ageostrophic

ERA-Interim

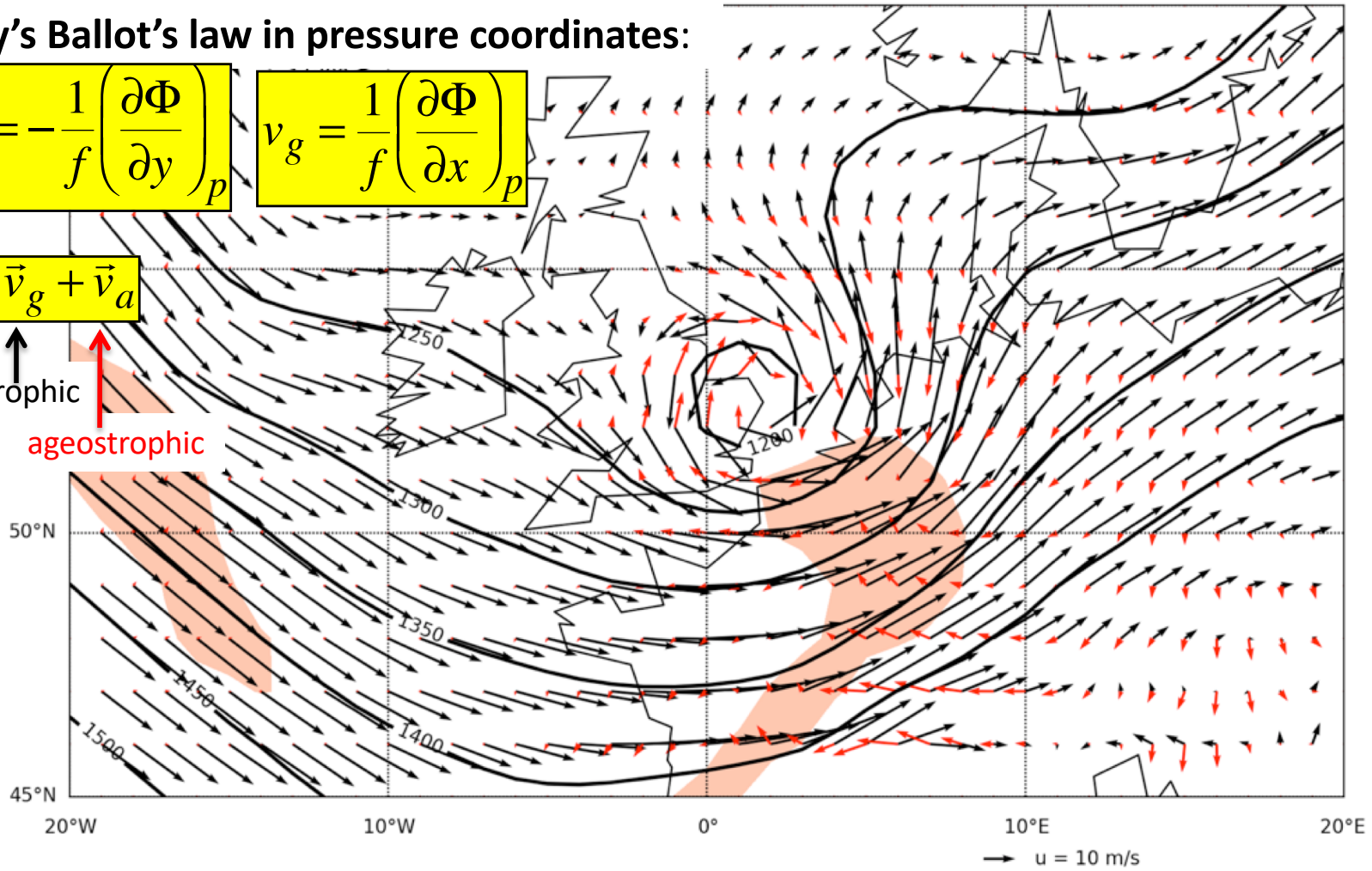


Figure 1.94: height at the 850 hPa constant pressure level (black contours, labeled in metres), the **geostrophic wind vector, (u_g,v_g)** (**black arrows**), the **ageostrophic wind vector** (**red arrows**) and the absolute wind velocity (red shading, starting at 25 m/s) on 3 March 1995, 00 UTC. Departures from geostrophic balance are appreciable only in the cyclonic circulation and in the vicinity of the Alps

section 1.31

Quasi-geostrophic approximation

(1) The wind velocity is equal to the sum of the balanced ("**geostrophic**") wind (subscript ***g***) and the deviation from this balanced wind (subscript ***a***, standing for "**ageostrophic**):

$$\vec{v} = \vec{v}_g + \vec{v}_a \qquad |\vec{v}_a| << |\vec{v}_g|$$

(2) The **Coriolis parameter** is a linear function of latitude: $f = f_0 + \beta(y - y_0)$

where f_0 is the value of f at a chosen reference latitude, $y=y_0$ (usually 45°N or S), and where the so-called "***β*-parameter**" is defined by

$$\beta \equiv \frac{df}{dy} \text{ at } y = y_0$$

$$f_0 >> \beta(y - y_0)$$

(3) The quasi-geostrophic approximation is usually applied to the equations in pressure coordinates. In the balanced state the material derivative and the curvature terms in equations of motion (1.191a,b) are neglected, so that the **balanced ("geostrophic") wind** is approximated by

$$u_g \approx -\frac{1}{f_0}\frac{\partial \Phi}{\partial y} \qquad v_g \approx \frac{1}{f_0}\frac{\partial \Phi}{\partial x}$$

section 1.31

Quasi-geostrophic (QG) vorticity

1. Geostrophic wind according to the quasi-geostrophic approximation:

$$u_g \approx -\frac{1}{f_0}\frac{\partial \Phi}{\partial y} \qquad v_g \approx \frac{1}{f_0}\frac{\partial \Phi}{\partial x} \qquad \textbf{(1.246)}$$

f **assumed constant!**

2. Quasi-geostrophic vorticity, ζ_g, can be expressed in terms of the **geopotential**, using (1.246), as

$$\zeta_g \equiv \frac{\partial v_g}{\partial x} - \frac{\partial u_g}{\partial y} = \frac{1}{f_0}\left(\frac{\partial^2 \Phi}{\partial x^2} + \frac{\partial^2 \Phi}{\partial y^2}\right) \equiv \frac{1}{f_0}\nabla_h^2 \Phi \qquad \textbf{(1.249)}$$

3. The **geopotential**, Φ, is determined by **"inverting"** the **Poisson (elliptic) equation**:

$$\nabla_h^2 \Phi = f_0 \zeta_g \qquad \textbf{(1.253)}$$

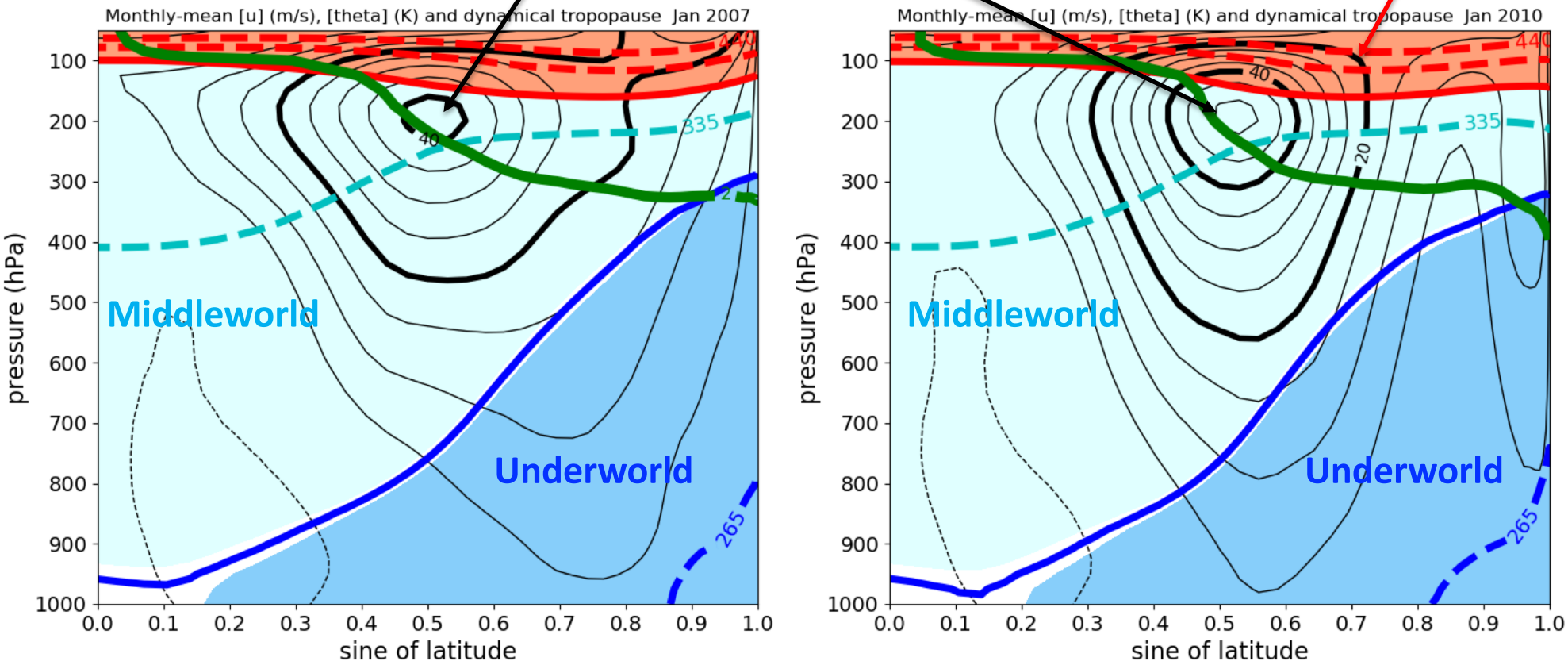
The **QG-approximation is the basis of a theory of planetary waves and cyclogenesis… later this lecture**

Zonal-mean zonal wind velocity: subtropical jet in January

2007

2010

Overworld

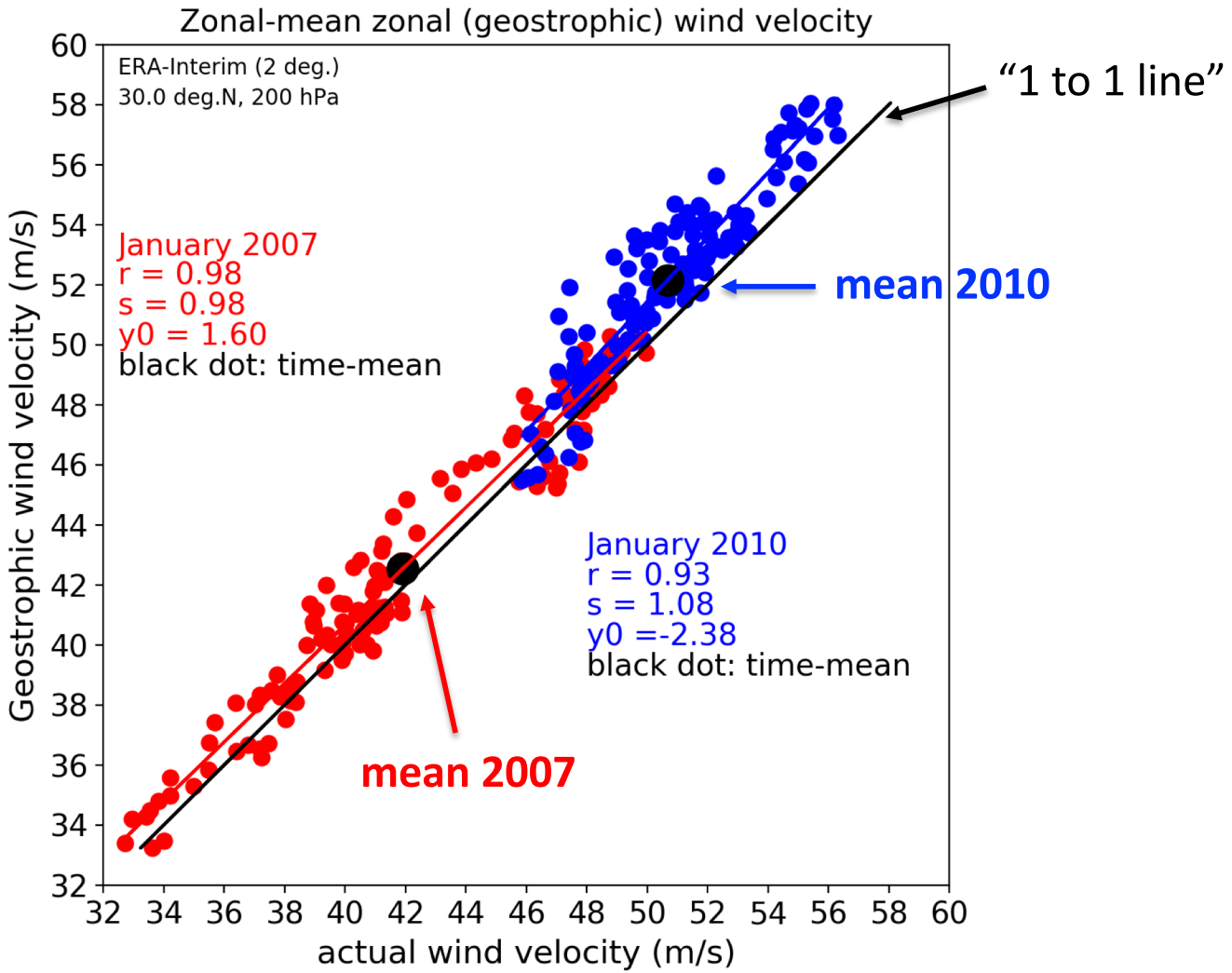


Black contours: [*u*], interval: 5 m/s, dashed negative, zero-contour not drawn.
Coloured contours: isentropes, labeled in K

ERAI_u_theta_zm_tm_nhJan2007

Geostrophic balance of the zonal-mean subtropical jet in January

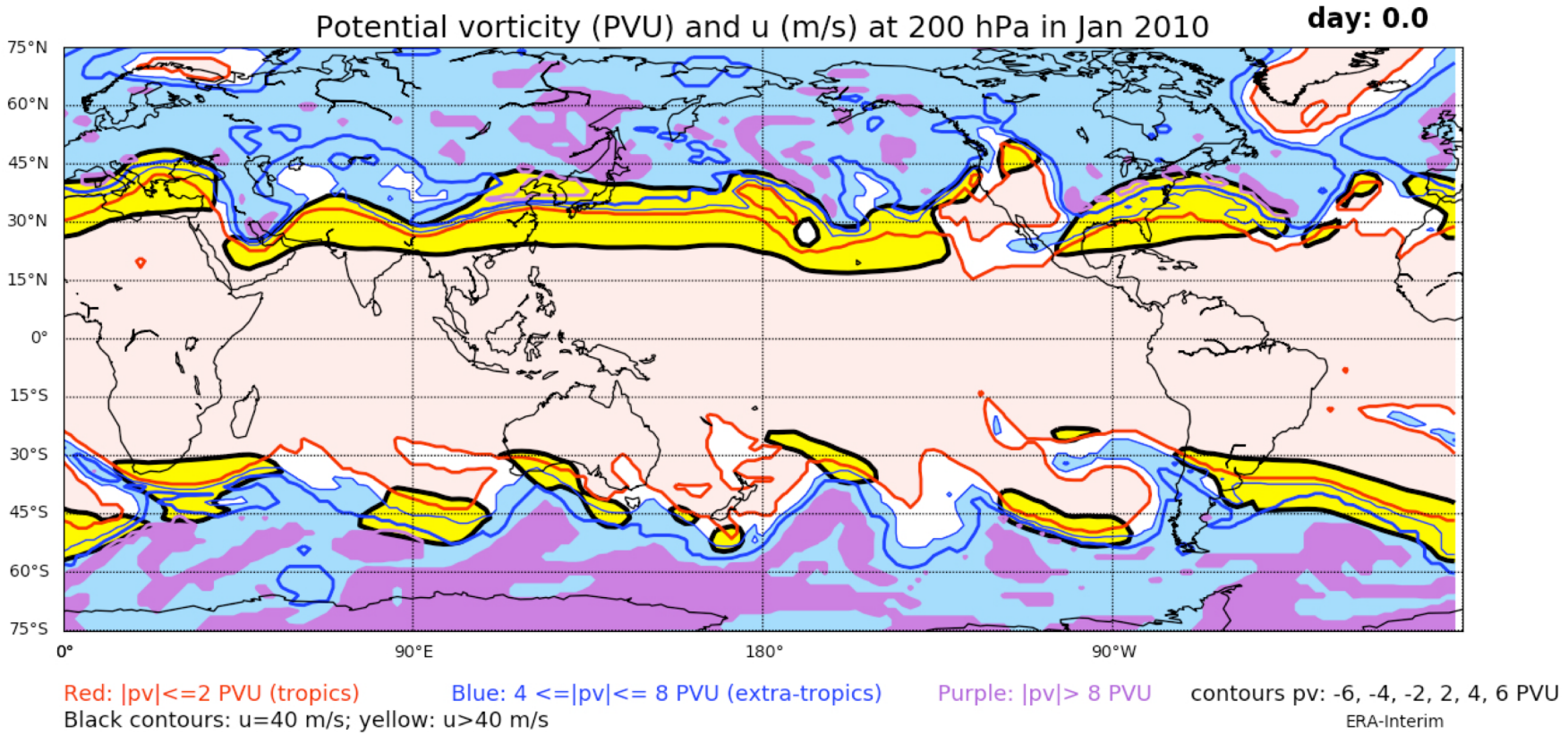
At **30°N** and **200 hPa** (see previous slide), based on 6-hourly data (124 points per month)



Plot-ERAI_geobalance_p.py

Figure_GeoWindvsActualWind

The zonal-mean subtropical jet and potential vorticity at 200 hPa in January 2010



Yellow shading represents the subtropical jet stream. Waves propagate along the jet. These waves are called **planetary waves**, or **Rossby-waves**…

STJ_pv_u200hPaJan2010.mov

Carl Gustav Rossby (1898-1957)

Time-cover of December 1956

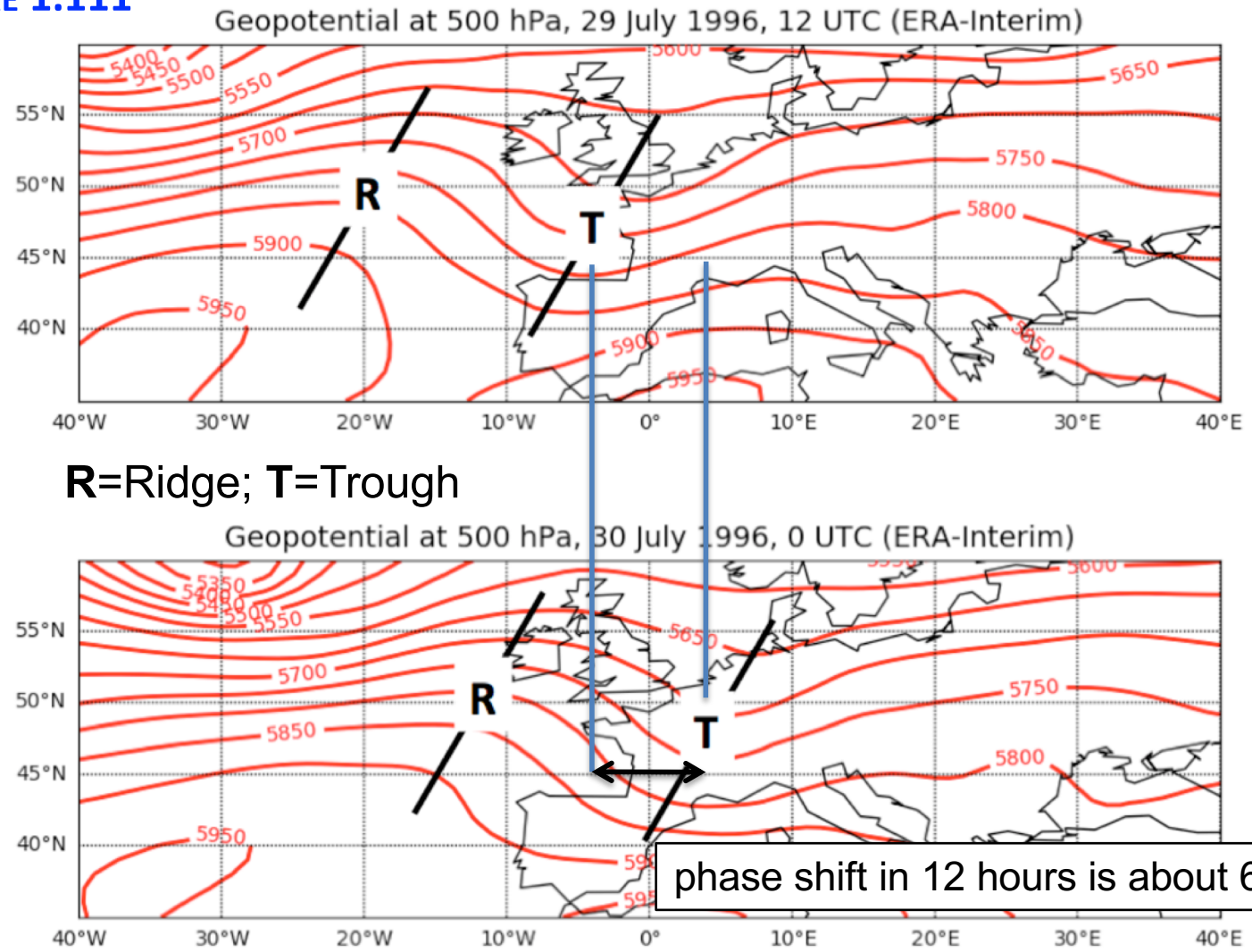
In the 1930's Rossby identified planetary waves and constructed a simplified theory of the mechanism and propagation of planetary waves.



Section 1.37

Planetary (Rossby) waves

FIGURE 1.111



The phase speed in x-direction is about +16 m/s

Quasi-geostrophic equations of motion

Slide 305: $f = f_0 + \beta(y - y_0)$

$$v_g \approx \frac{1}{f_0}\frac{\partial \Phi}{\partial x}$$

$$u_g \approx -\frac{1}{f_0}\frac{\partial \Phi}{\partial y}$$

Equation of motion in pressure coordinates:

$$\frac{d_g u_g}{dt} \approx -\frac{\partial \Phi}{\partial x} + (f_0 + \beta(y - y_0))(v_g + v_a)$$

Advection with the geostrophic wind:

$$\frac{\partial u_g}{\partial t} + u_g\left(\frac{\partial u_g}{\partial x}\right)_p + v_g\left(\frac{\partial u_g}{\partial y}\right)_p = f_0 v_a + \beta(y - y_0)v_g$$

(in pressure coordinates)

Quasi-geostrophic equations of motion

Equations of motion:

$$\frac{\partial u_g}{\partial t} + u_g\left(\frac{\partial u_g}{\partial x}\right)_p + v_g\left(\frac{\partial u_g}{\partial y}\right)_p = f_0 v_a + \beta(y - y_0)v_g$$

$$\frac{\partial v_g}{\partial t} + u_g\left(\frac{\partial v_g}{\partial x}\right)_p + v_g\left(\frac{\partial v_g}{\partial y}\right)_p = -f_0 v_a - \beta(y - y_0)u_g$$

Geostrophic vorticity

$$\zeta_g \equiv \frac{\partial v_g}{\partial x} - \frac{\partial u_g}{\partial y} \approx \frac{1}{f_0}\nabla^2\Phi$$

Quasi-geostrophic equations of motion

Equations of motion:

$$\frac{\partial u_g}{\partial t} + u_g\left(\frac{\partial u_g}{\partial x}\right)_p + v_g\left(\frac{\partial u_g}{\partial y}\right)_p = f_0 v_a + \beta(y - y_0)v_g \qquad (1.247a)$$

$$\frac{\partial v_g}{\partial t} + u_g\left(\frac{\partial v_g}{\partial x}\right)_p + v_g\left(\frac{\partial v_g}{\partial y}\right)_p = -f_0 v_a - \beta(y - y_0)u_g \qquad (1.247b)$$

Geostrophic vorticity

$$\zeta_g \equiv \frac{\partial v_g}{\partial x} - \frac{\partial u_g}{\partial y} \approx \frac{1}{f_0}\nabla^2\Phi$$

Quasi-geostrophic vorticity eqn.

$$\boxed{\frac{\partial \zeta_g}{\partial t} + u_g\left(\frac{\partial \zeta_g}{\partial x}\right)_p + v_g\left(\frac{\partial \zeta_g}{\partial y}\right)_p = -f_0\left(\frac{\partial u_a}{\partial x} + \frac{\partial v_a}{\partial y}\right) - \beta v_g} \qquad (1.250)$$

Quasi-geostrophic equations of motion

Equations of motion:

$$\frac{\partial u_g}{\partial t} + u_g\left(\frac{\partial u_g}{\partial x}\right)_p + v_g\left(\frac{\partial u_g}{\partial y}\right)_p = f_0 v_a + \beta(y - y_0)v_g$$

$$\frac{\partial v_g}{\partial t} + u_g\left(\frac{\partial v_g}{\partial x}\right)_p + v_g\left(\frac{\partial v_g}{\partial y}\right)_p = -f_0 v_a - \beta(y - y_0)u_g$$

Geostrophic vorticity

$$\zeta_g \equiv \frac{\partial v_g}{\partial x} - \frac{\partial u_g}{\partial y} \approx \frac{1}{f_0}\nabla^2\Phi$$

Quasi-geostrophic vorticity eqn.

$$\boxed{\frac{\partial \zeta_g}{\partial t} + u_g\left(\frac{\partial \zeta_g}{\partial x}\right)_p + v_g\left(\frac{\partial \zeta_g}{\partial y}\right)_p = -f_0\left(\frac{\partial u_a}{\partial x} + \frac{\partial v_a}{\partial y}\right) - \beta v_g}$$

With:

$$\zeta_g = \frac{1}{f_0}\nabla^2\Phi$$

$$\boxed{\frac{\partial \nabla^2\Phi}{\partial t} + u_g\left(\frac{\partial \nabla^2\Phi}{\partial x}\right)_p + v_g\left(\frac{\partial \nabla^2\Phi}{\partial y}\right)_p = -f_0^2\left(\frac{\partial u_a}{\partial x} + \frac{\partial v_a}{\partial y}\right) - \beta\frac{\partial \Phi}{\partial x}}$$

Section 1.31

Quasi-geostrophic vorticity equation

Previous slide:

$$\frac{\partial \nabla^2 \Phi}{\partial t} + u_g \left(\frac{\partial \nabla^2 \Phi}{\partial x} \right)_p + v_g \left(\frac{\partial \nabla^2 \Phi}{\partial y} \right)_p = -f_0^2 \left(\frac{\partial u_a}{\partial x} + \frac{\partial v_a}{\partial y} \right) - \beta \frac{\partial \Phi}{\partial x} \qquad (1.251)$$

Barotropic vorticity equation: $\frac{\partial u_a}{\partial x} + \frac{\partial v_a}{\partial y} = 0$

$$\frac{\partial \nabla^2 \Phi}{\partial t} + u_g \left(\frac{\partial \nabla^2 \Phi}{\partial x} \right)_p + v_g \left(\frac{\partial \nabla^2 \Phi}{\partial y} \right)_p = -\beta \frac{\partial \Phi}{\partial x} \qquad (1.252)$$

$$u_g = -\frac{1}{f_0} \frac{\partial \Phi}{\partial y} \qquad v_g = \frac{1}{f_0} \frac{\partial \Phi}{\partial x} \qquad (1.246)$$

$$\frac{\partial \nabla^2 \Phi}{\partial t} - \frac{1}{f_0} \frac{\partial \Phi}{\partial y} \left(\frac{\partial \nabla^2 \Phi}{\partial x} \right)_p + \frac{1}{f_0} \frac{\partial \Phi}{\partial x} \left(\frac{\partial \nabla^2 \Phi}{\partial y} \right)_p = -\beta \frac{\partial \Phi}{\partial x}$$

One unknown variable: Φ

Section 1.37

A simple theory of planetary waves

The "**barotropic" quasi-geostrophic vorticity equation**:

$$\frac{\partial \nabla^2 \Phi}{\partial t} + u_g \left(\frac{\partial \nabla^2 \Phi}{\partial x} \right)_p + v_g \left(\frac{\partial \nabla^2 \Phi}{\partial y} \right)_p = -\beta \frac{\partial \Phi}{\partial x}$$

Assume a basic state zonal current, $[u]$, is **in geostrophic balance** with the zonal mean geopotential, i.e.

$$[u] \equiv U_g = -\frac{1}{f_0} \left(\frac{\partial [\Phi]}{\partial y} \right)_p$$

where $\Phi(x,y,p,t) = [\Phi] + \Phi^*(x,y,p,t)$

perturbation of zonal mean state (waves)

Assume: $|u^*| << |U_g|$ and $|\Phi^*| << |[\Phi]|$

Approximate vorticity equation: (linearised)

$$\frac{\partial \nabla_h^2 (\Phi^*)}{\partial t} + U_g \frac{\partial \nabla_h^2 (\Phi^*)}{\partial x} + \beta \frac{\partial \Phi^*}{\partial x} \approx 0$$

Section 1.37

Wave-like solutions of linearized vorticity equation

Previous slide: $$\frac{\partial \nabla_h^2(\Phi^*)}{\partial t} + U_g \frac{\partial \nabla_h^2(\Phi^*)}{\partial x} + \beta \frac{\partial \Phi^*}{\partial x} \approx 0$$

On an infinite "beta-plane" (β is constant), insert wave-like solution:

$$\Phi^* = \mathrm{Re}\{A \exp[i(lx + my - \omega t)]\}$$

$$\omega \equiv \frac{2\pi}{T};\ l \equiv \frac{2\pi}{L_x};\ m \equiv \frac{2\pi}{L_y}$$

frequency

period

wave number

wavelength

Result is a **dispersion relation**: $$\omega = U_g l - \frac{\beta l}{l^2 + m^2}$$

Section 1.37

Dispersion relation for barotropic planetary (Rossby) waves

Dispersion relation: $$\omega = U_g l - \frac{\beta l}{l^2 + m^2} \qquad (1.312)$$

Phase speed: $$c_x = \frac{\omega}{l} = U_g - \frac{\beta}{l^2 + m^2} \qquad (1.313)$$

If U_g=0, phase speed of Rossby waves is negative (westward propagation).
Rossby waves are dispersive, i.e. phase speed depends on wavelength.

The quasi-geostropic approximation "filters" all waves except Rossby waves!

such as sound waves and buoyancy waves

Section 1.37

Dispersion relation for barotropic planetary (Rossby) waves

Dispersion relation: $$\omega = U_g l - \frac{\beta l}{l^2 + m^2} \qquad (1.312)$$

Phase speed: $$c_x = \frac{\omega}{l} = U_g - \frac{\beta}{l^2 + m^2} \qquad (1.313)$$

If U_g=0, phase speed of Rossby waves is negative (westward propagation). Rossby waves are dispersive, i.e. phase speed depends on wavelength.

Group speed:

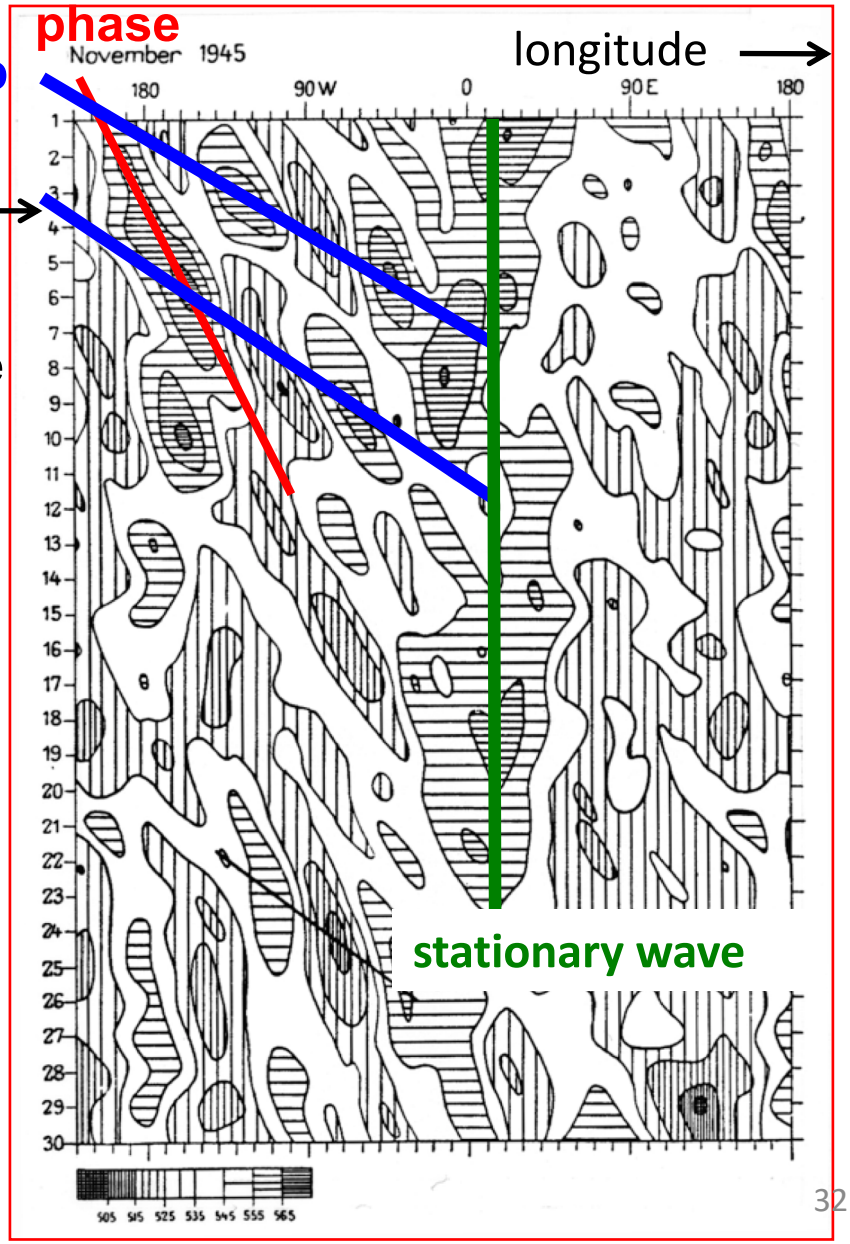
$$c_{gx} = \frac{\partial \omega}{\partial l} = U_g - \frac{\beta}{l^2 + m^2} + 2\beta l^2 \left(l^2 + m^2\right)^{-2} = c_x + 2\beta l^2 \left(l^2 + m^2\right)^{-2}$$

Section 1.37

"Hovmöller" diagram

FIGURE 1.112

November 1945



Previous slide:

$$c_x = \frac{\omega}{l} = U_g - \frac{\beta}{l^2 + m^2}$$

$$c_{gx} = c_x + 2\beta l^2 \left(l^2 + m^2\right)^{-2} > c_x$$

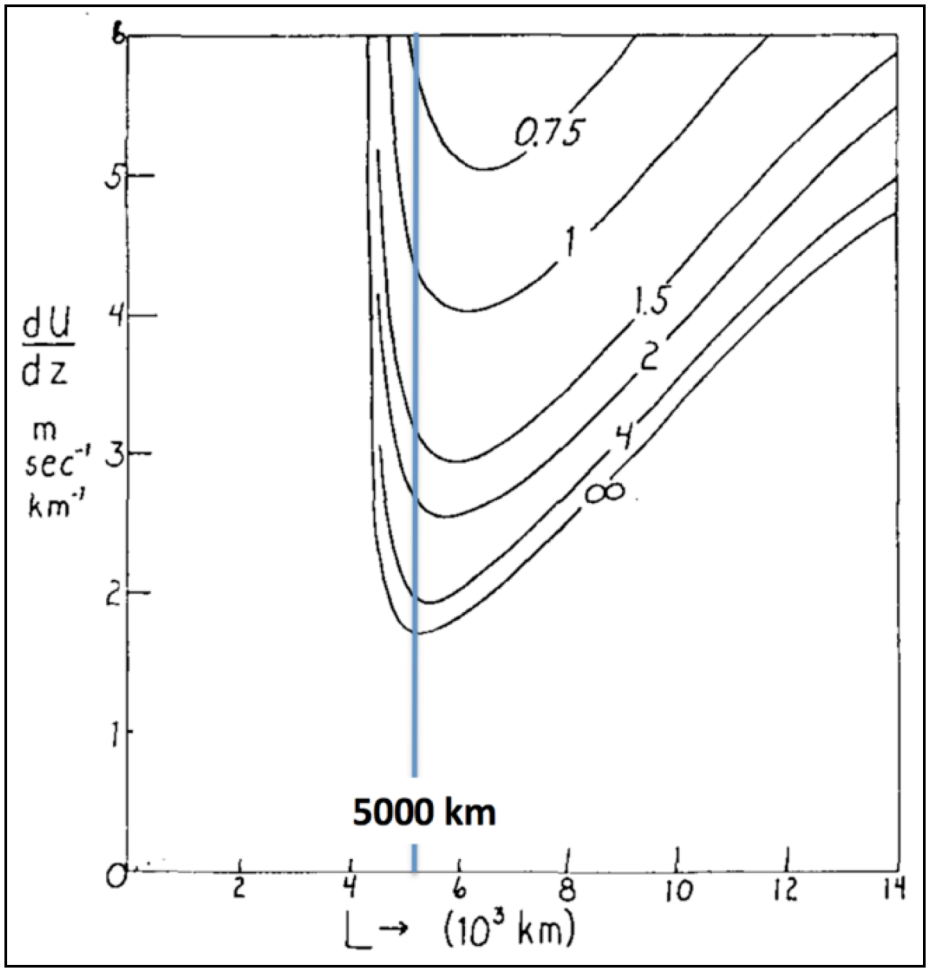
(page 252)

PROBLEM 1.34. <u>The basic state zonal flow</u> : Estimate the zonally averaged, time-mean value of U_g in the western hemisphere for the month of November 1945 from the figure on the right.

http://www.staff.science.uu.nl/~delde102/Hovmoller[1949].pdf

Section 1.38

Instability of a baroclinic Rossby waves



$$\Phi^* = \mathrm{Re}\{A\exp[i(lx+my-\omega t)]\}$$

Instability if $\omega^2<0$.

Only possible in a baroclinic atmosphere*

When is the atmosphere baroclinic ?

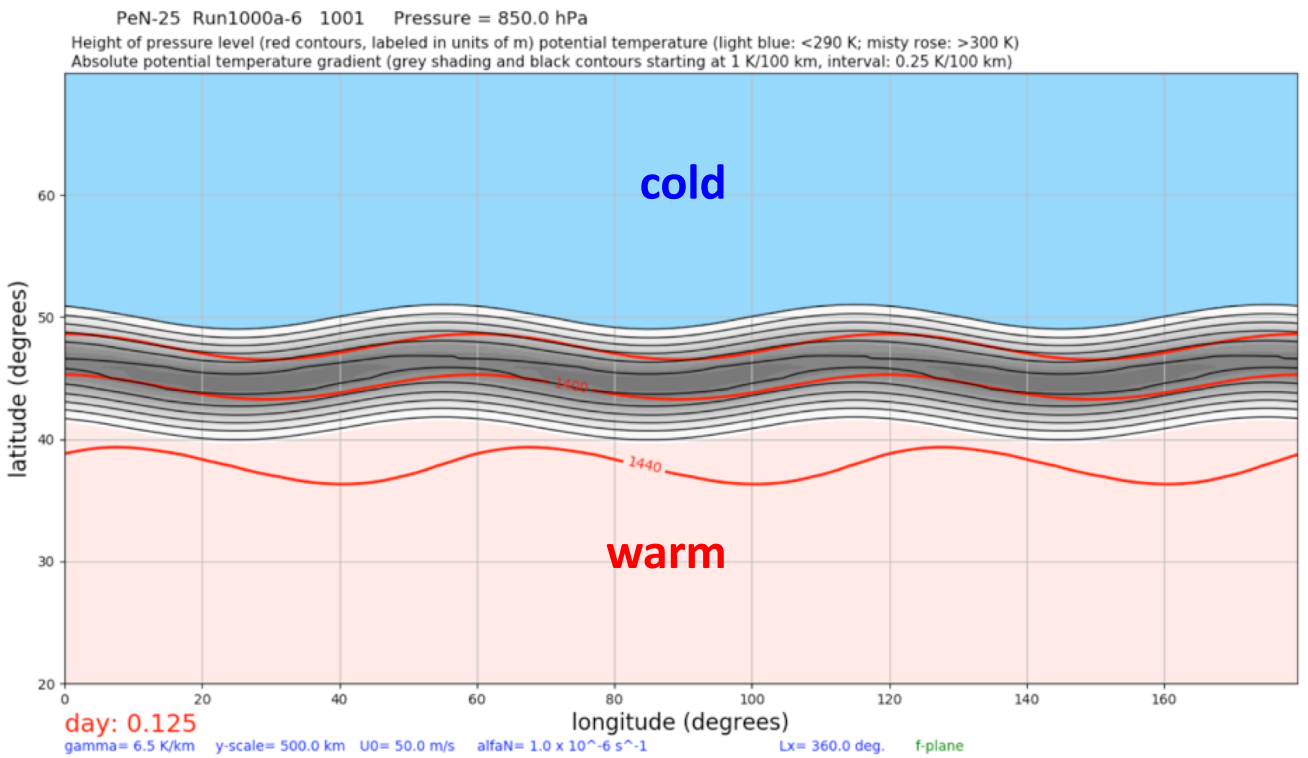
N.A. Phillips, 1954: Energy Transformations and meridional circulations associated with simple baroclinic waves in a two-level, quasi-geostrophic Model. **Tellus, 6** (3), 274-286.

FIGURE 1.113. Time, in days, required for an unstable wave to double its amplitude, given as a function of the vertical wind shear of the basic current and the wavelength (Phillips 1954).

*chapter 9, (https://webspace.science.uu.nl/~delde102/AtmosphericDynamics.htm)

Section 1.39

Instability of a baroclinic Rossby waves

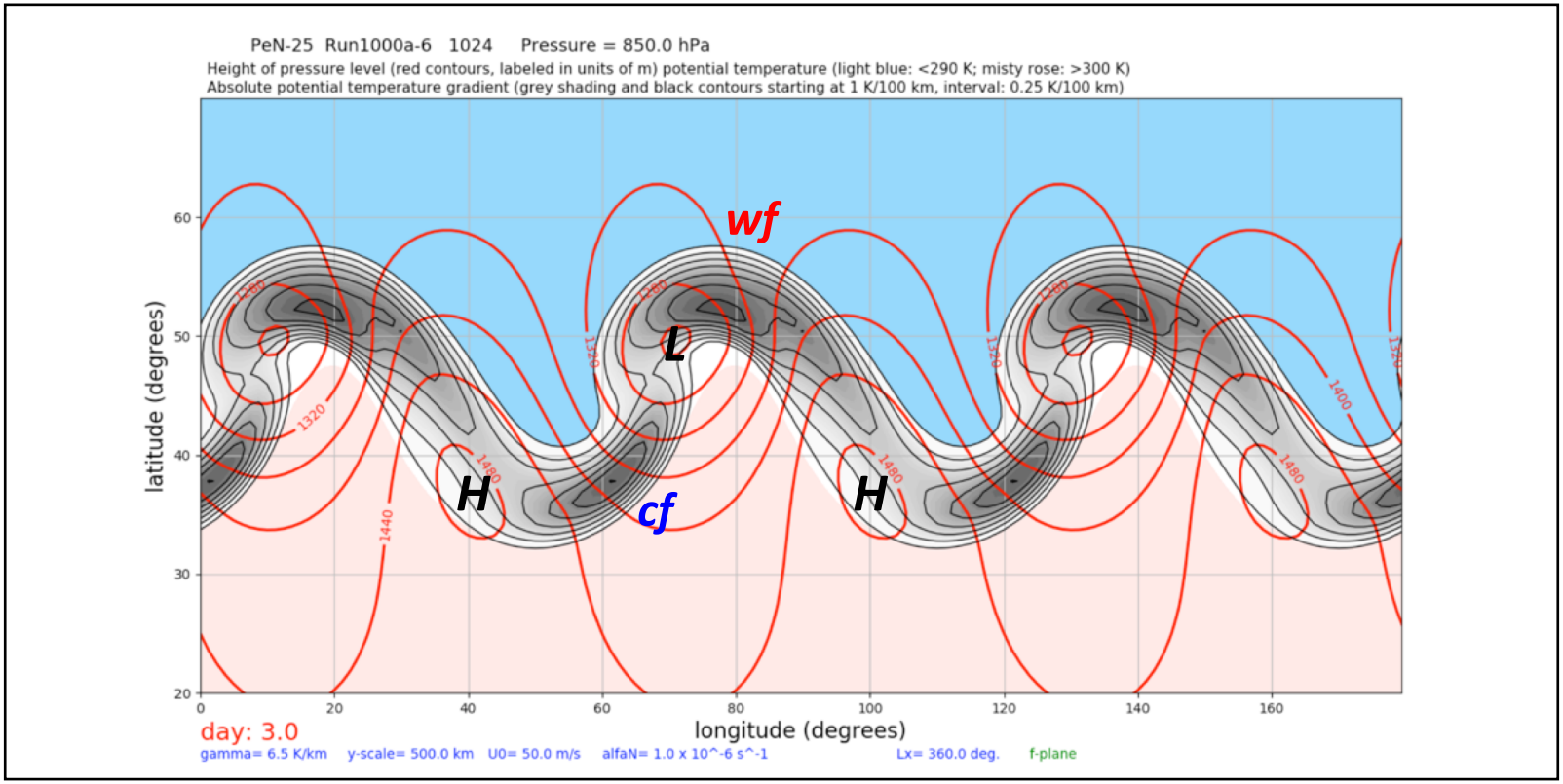


Initial condition: small-amplitude wave

FIGURE 1.117. Results of a numerical simulation of a baroclinic wave life cycle on the "*f*-plane". Half of the longitudinal extent of the domain of computation is shown. Shown are fields at 850 hPa, on day 0 (3 hours after initialisation), and on day 3 (next slide), of temperature (light blue: <290 K; misty rose: >300 K), height (red contours, labeled in m) and absolute isobaric temperature gradient (shading and black contours starting at 1 K per 100 km, contour interval is 0.25 K per 100 km). Surface pressure is assumed constant (1013 hPa) at initial time.

Section 1.39

Instability of a baroclinic Rossby waves



Day 3

The temperature gradients on day 3 in both the cold front (*cf*) and the warm front (*wf*) are larger than the maximum temperature gradient at initial time (previous slide)! An area characterised by low geopotential height (labeled ***L***) is observed in the region between the warm front and the cold front on day 3 of the simulation. The resolution of the model is 1.25° in longitude, 0.9 ° in latitude and 40 hPa in the vertical. This implies 288 x 91 x 25 grid points, as opposed to 16 x 17 x 2 grid points in the first simulation performed in the early 1950's by Norman Phillips (**next slide**). This simulation is analysed further in **Part I** (**chapter 1**). Further technical details of this simulation are given in **chapter 10** of the lecture notes (https://webspace.science.uu.nl/~delde102/AtmosphericDynamics.htm)

Animation: https://webspace.science.uu.nl/~delde102/Fronts850hPa_Run1000b-6days%20(Converted)

Section 1.39

Norwegian model of cyclone life cycle

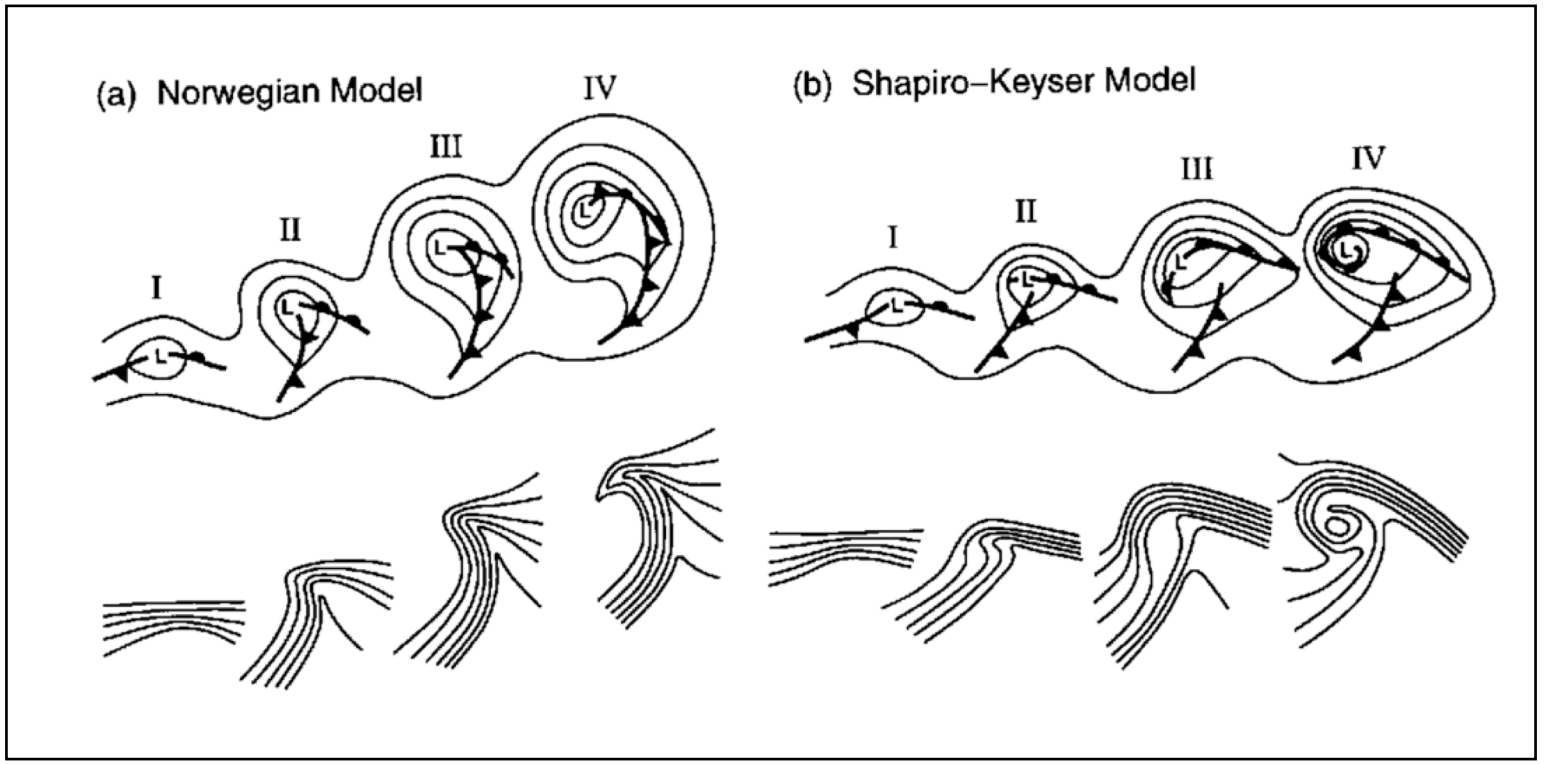


FIGURE 1.116. The life cycle of an **extratropical cyclone**, according to the Norwegian (Bergen) school (left) and according to M.A. Shapiro and D. Keyser (1990) (right). The life cycle according to Shapiro and Keyser is characterized by four stages: (I) incipient frontal cyclone; (II) frontal fracture; (III) bent-back warm front and frontal T-bone; (IV) warm-core frontal seclusion. Upper panels: sea level isobars and fronts. Lower panels: surface isotherms.

Norwegian model of a cyclone

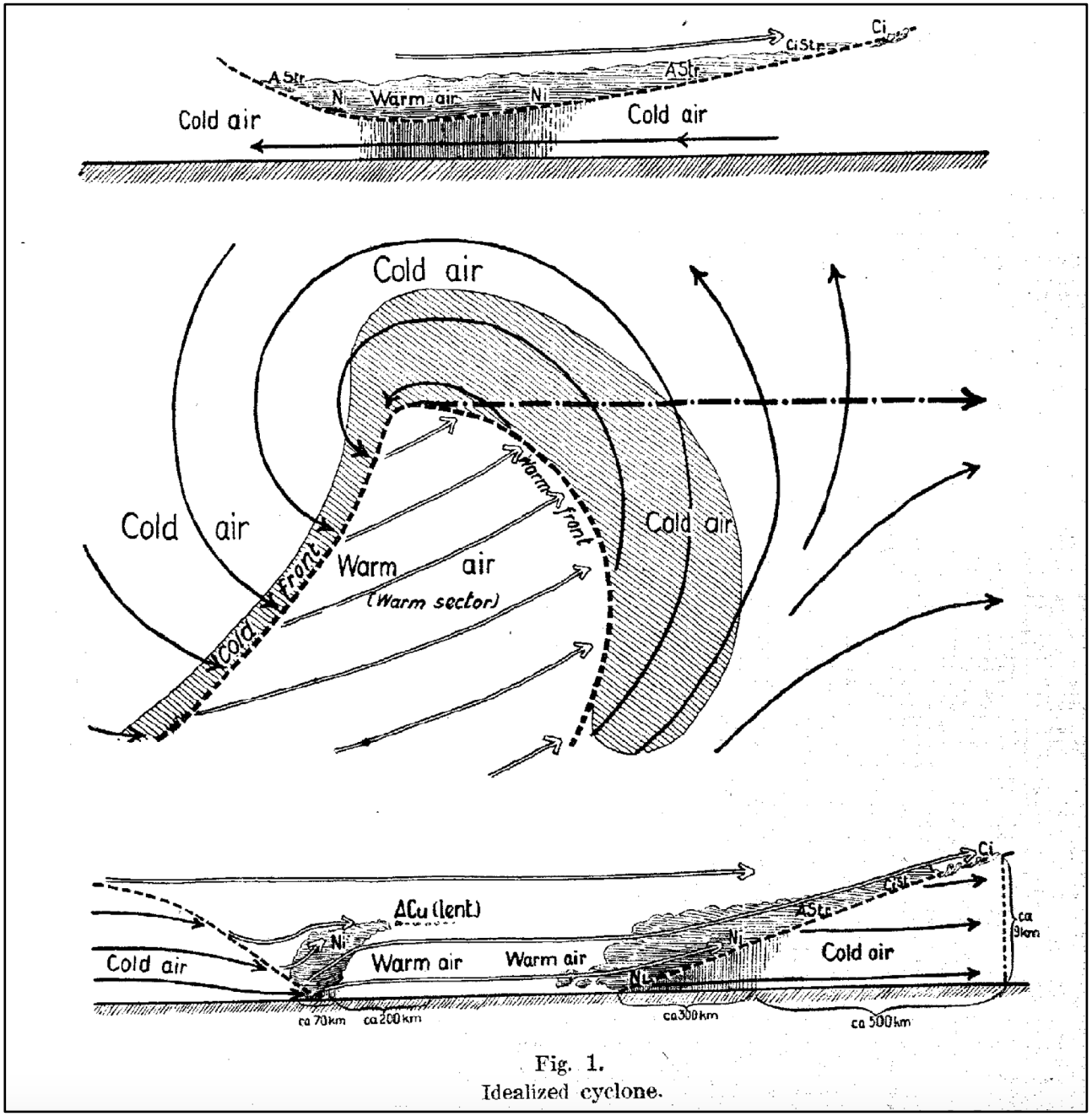


Fig. 1.
Idealized cyclone.

"Life cycle of cyclones and the polar front theory of atmospheric circulation"

Bjerknes and Solberg, 1922

Section 1.39

Norwegian model of cyclone

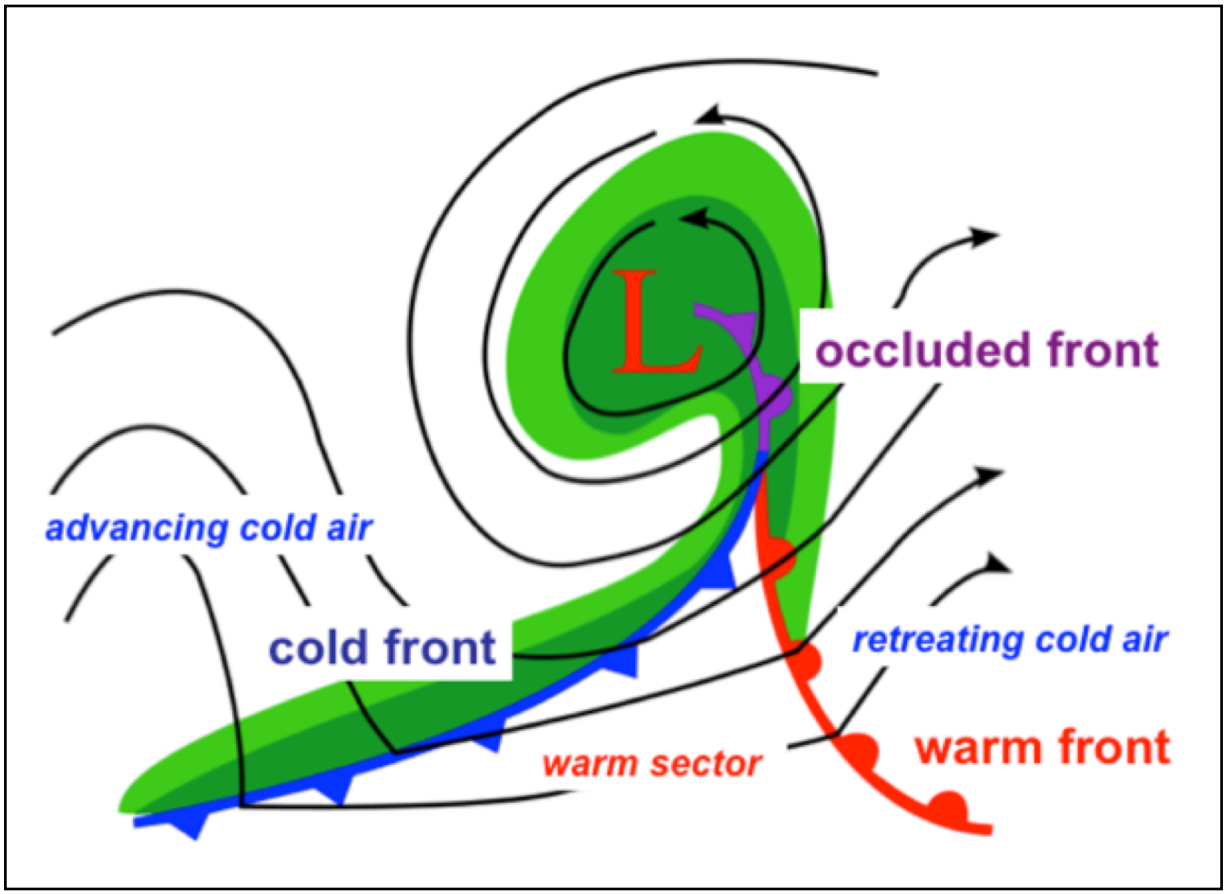


FIGURE 1.115. Conceptual model of the positions of fronts near the Earth's surface within a full-grown middle latitude cyclone. This conceptual model is due to the *Bergen school*.

Next:

Dynamic Meteorology: lecture 12

11/1/2023 (Wednesday, 13:15)

Sections 1.32 and 1.33 (part G)

Balanced circular vortex (gradient wind balance)
Example: tropical cyclone (section 1.32)
Potential vorticity inversion (section 1.33)
Insights from PV-inversion on the thermodynamic structure of balanced flow
Kleinschmidt's principles
The ageostrophic wind, accelerations of the jestream, jet-streaks

Tutorial after this lecture (11)
Project 5
1.20, 1.21 (part E) p. 153
1.22 (part F) p. 189
Problems 1.34 and 1.35 (part G) p. 252-253

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(http://www.uu.nl/~nvdelden/dynmeteorology.htm)