

Dynamic Meteorology: lecture 15

21/1/2022 (Friday, 9:00-10:45)

Sections 1.42-1.43 (part I)

Vertical motion in statically stable atmosphere part 2

Ageostrophic motion is due to frontogenesis

Large-scale vertical motion: the omega-equation

Relation between Q-vector and large-scale vertical motion

Case of problem 1.35 : vertical velocity according to the **omega equation**

After this lecture: eight presentations of project 5

Abhi, Filippo, Maren and Martin - NAO, EA, temp Florida, temp Middle East

Jim, Koen and Derk – Tropical cyclones and NAO/NAM

Famke, Marijn and Hannah – AA, jet stream shift and weakening of storm tracks

Laura, Elia and Emilia – cyclones, SST, wind and moisture

Rens, Marjolein and Marius - hurricanes, SST, jet and rising CO2 levels

Jos and Ben - Sea Ice, NAO, SOI, Winter temp NL

Anja, Rainette, Joao and Mattia – Sea ice in arctic and Antarctic, (ant) arctic oscillations and global mean temp

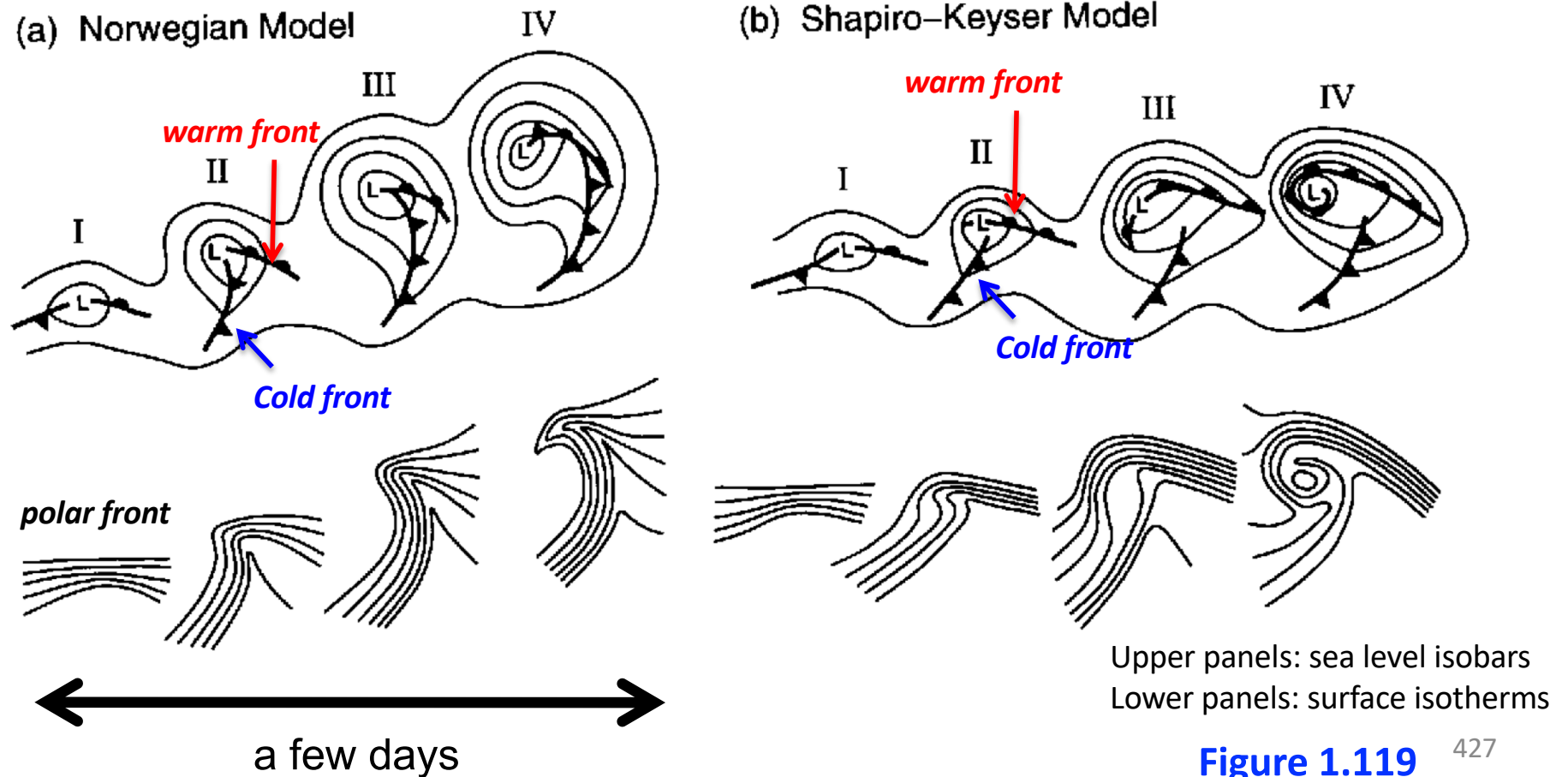
Lorenzo, Fiona and Lotje – Himalayan ice thickness, rising CO2, rising temp and precipitation

(a.j.vandelden@uu.nl)

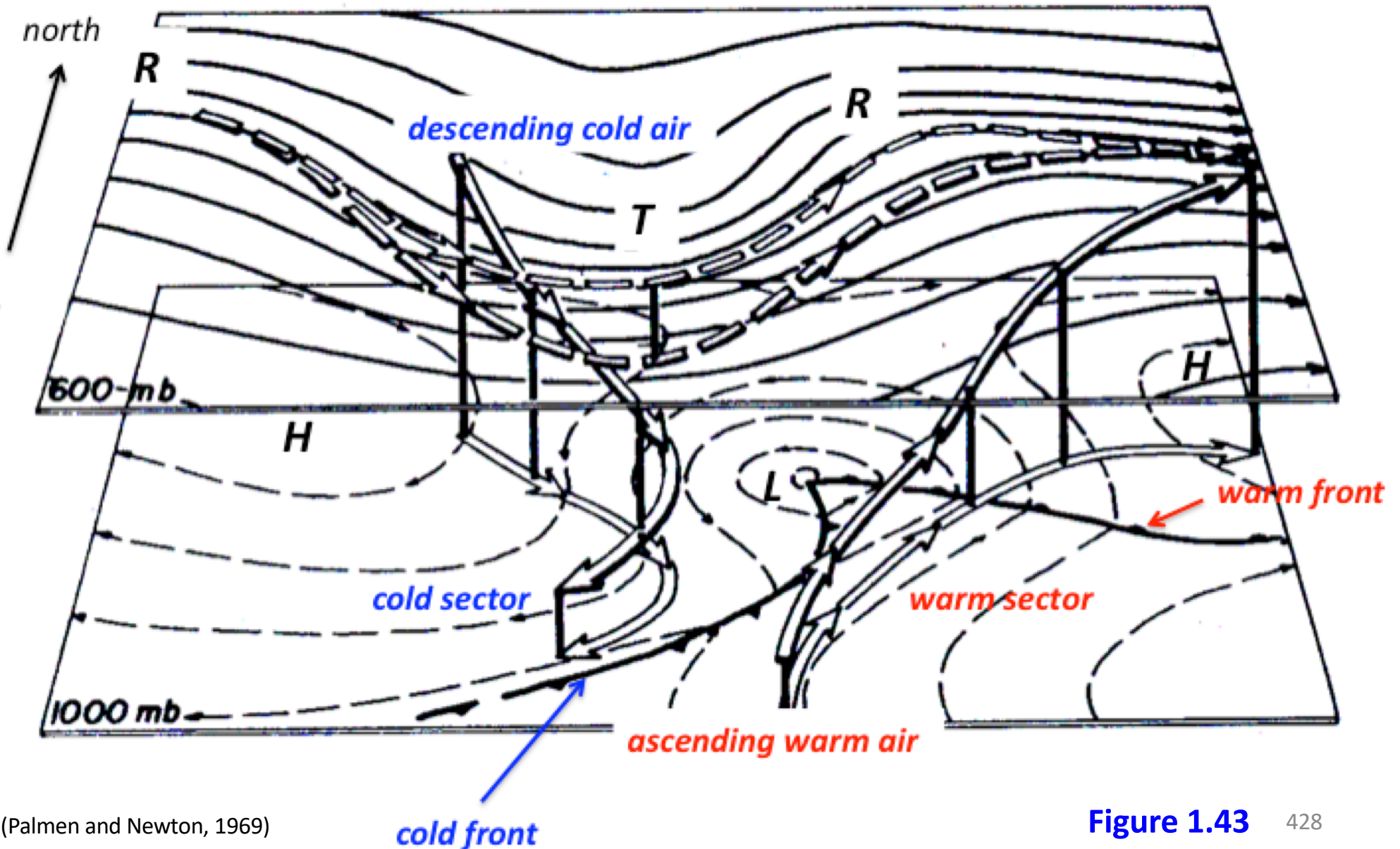
(<http://www.uu.nl/~nvdelden/dynmeteorology.htm>)

Conceptual model of life-cycle of unstable baroclinic wave

Starts with a “wave” in the “Polar Front”. This wave grows due to the instability of thermal wind balance, producing “secondary fronts”.



Ascent in warm sector and descent in cold sector of a planetary wave



Equations needed to study the dynamics of frontogenesis in a quasi-geostrophic atmosphere

Thermal wind balance: $\frac{\partial u_g}{\partial p} = \frac{R}{pf_0} \frac{\partial T}{\partial y}$ $\frac{\partial v_g}{\partial p} = -\frac{R}{pf_0} \frac{\partial T}{\partial x}$ eq.1.347

and $\frac{\partial T}{\partial t} + u_g \frac{\partial T}{\partial x} + v_g \frac{\partial T}{\partial y} - S_p \omega = + \frac{J}{c_p}$ (previous lecture)

also: $\frac{\partial u_g}{\partial t} + u_g \frac{\partial u_g}{\partial x} + v_g \frac{\partial u_g}{\partial y} = f_0 v_a + \beta y v_g$ (1.247a)

This lecture investigates the relation between vertical motion in a statically stable atmosphere and frontogenesis based upon the above equations.

Equations needed to study the dynamics of frontogenesis in a quasi-geostrophic atmosphere

Thermal wind balance: $\frac{\partial u_g}{\partial p} = \frac{R}{pf_0} \frac{\partial T}{\partial y}$ $\frac{\partial v_g}{\partial p} = -\frac{R}{pf_0} \frac{\partial T}{\partial x}$ eq.1.347

and $\frac{\partial T}{\partial t} + u_g \frac{\partial T}{\partial x} + v_g \frac{\partial T}{\partial y} - S_p \omega = + \frac{J}{c_p}$ (previous lecture)

also: $\frac{\partial u_g}{\partial t} + u_g \frac{\partial u_g}{\partial x} + v_g \frac{\partial u_g}{\partial y} = f_0 v_a + \beta y v_g$ (1.247a)

Start by neglecting ageostrophic motion ($\omega=0$), and assume $\beta=0$ and $J=0$:

then $\frac{\partial T}{\partial t} + u_g \frac{\partial T}{\partial x} + v_g \frac{\partial T}{\partial y} = 0$

becomes $\frac{\partial u_g}{\partial t} + u_g \frac{\partial u_g}{\partial x} + v_g \frac{\partial u_g}{\partial y} = 0$

These assumptions are not realistic but “didactic”, as we shall see later

Dynamical consequences of frontogenesis in a quasi-geostrophic atmosphere

Thermal wind balance:

$$\frac{\partial u_g}{\partial p} = \frac{R}{pf_0} \frac{\partial T}{\partial y} \quad \frac{\partial v_g}{\partial p} = -\frac{R}{pf_0} \frac{\partial T}{\partial x}$$

Previous slide:

$$\frac{\partial T}{\partial t} + u_g \frac{\partial T}{\partial x} + v_g \frac{\partial T}{\partial y} = 0$$

$$\frac{\partial u_g}{\partial t} + u_g \frac{\partial u_g}{\partial x} + v_g \frac{\partial u_g}{\partial y} = 0$$

Do the derivations

$$\frac{d_g}{dt} \left(\frac{\partial T}{\partial y} \right) = -\frac{\partial v_g}{\partial y} \frac{\partial T}{\partial y} - \frac{\partial u_g}{\partial y} \frac{\partial T}{\partial x}$$

$$\frac{d_g}{dt} \left(\frac{\partial u_g}{\partial p} \right) = -\frac{\partial u_g}{\partial p} \frac{\partial u_g}{\partial x} - \frac{\partial v_g}{\partial p} \frac{\partial u_g}{\partial y}$$

Are these equations consistent with thermal wind balance?

Dynamical consequences of frontogenesis in a quasi-geostrophic atmosphere

Thermal wind balance:

$$\frac{\partial u_g}{\partial p} = \frac{R}{pf_0} \frac{\partial T}{\partial y}$$

$$\frac{\partial v_g}{\partial p} = -\frac{R}{pf_0} \frac{\partial T}{\partial x}$$

$$\frac{\partial T}{\partial t} + u_g \frac{\partial T}{\partial x} + v_g \frac{\partial T}{\partial y} = 0$$

$$\frac{\partial u_g}{\partial t} + u_g \frac{\partial u_g}{\partial x} + v_g \frac{\partial u_g}{\partial y} = 0$$

$$\frac{d_g}{dt} \left(\frac{\partial T}{\partial y} \right) = -\frac{\partial v_g}{\partial y} \frac{\partial T}{\partial y} - \frac{\partial u_g}{\partial y} \frac{\partial T}{\partial x}$$

$$\frac{d_g}{dt} \left(\frac{\partial u_g}{\partial p} \right) = -\frac{\partial u_g}{\partial p} \frac{\partial u_g}{\partial x} - \frac{\partial v_g}{\partial p} \frac{\partial u_g}{\partial y} = \frac{R}{f_0 p} \left(-\frac{\partial u_g}{\partial x} \frac{\partial T}{\partial y} + \frac{\partial u_g}{\partial y} \frac{\partial T}{\partial x} \right)$$

$$\frac{\partial u_g}{\partial x} + \frac{\partial v_g}{\partial y} = 0$$

Dynamical consequences of frontogenesis in a quasi-geostrophic atmosphere

Thermal wind balance:

$$\frac{\partial u_g}{\partial p} = \frac{R}{pf_0} \frac{\partial T}{\partial y} \quad \frac{\partial v_g}{\partial p} = -\frac{R}{pf_0} \frac{\partial T}{\partial x}$$

$$\frac{\partial T}{\partial t} + u_g \frac{\partial T}{\partial x} + v_g \frac{\partial T}{\partial y} = 0$$

$$\frac{\partial u_g}{\partial t} + u_g \frac{\partial u_g}{\partial x} + v_g \frac{\partial u_g}{\partial y} = 0$$



$$\frac{d_g}{dt} \left(\frac{\partial T}{\partial y} \right) = -\frac{\partial v_g}{\partial y} \frac{\partial T}{\partial y} - \frac{\partial u_g}{\partial y} \frac{\partial T}{\partial x}$$

Subtract:

$$\frac{f_0 p}{R} \frac{d_g}{dt} \left(\frac{\partial u_g}{\partial p} \right) = \left(\frac{\partial v_g}{\partial y} \frac{\partial T}{\partial y} + \frac{\partial u_g}{\partial y} \frac{\partial T}{\partial x} \right)$$

Result:

$$\frac{d_g}{dt} \left(\frac{\partial T}{\partial y} \right) - \frac{f_0 p}{R} \frac{d_g}{dt} \left(\frac{\partial u_g}{\partial p} \right) = -2 \left(\frac{\partial v_g}{\partial y} \frac{\partial T}{\partial y} + \frac{\partial u_g}{\partial y} \frac{\partial T}{\partial x} \right)$$

→ next slide

Dynamical consequences of frontogenesis in a quasi-geostrophic atmosphere

Thermal wind balance:

$$\frac{\partial u_g}{\partial p} = \frac{R}{pf_0} \frac{\partial T}{\partial y} \quad \frac{\partial v_g}{\partial p} = -\frac{R}{pf_0} \frac{\partial T}{\partial x}$$

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$$\frac{d_g}{dt} \left(\frac{\partial T}{\partial y} \right) - \frac{f_0 p}{R} \frac{d_g}{dt} \left(\frac{\partial u_g}{\partial p} \right) = -2 \left(\frac{\partial v_g}{\partial y} \frac{\partial T}{\partial y} + \frac{\partial u_g}{\partial y} \frac{\partial T}{\partial x} \right)$$

$$\frac{d_g}{dt} \left(\frac{\partial T}{\partial y} - \frac{f_0 p}{R} \frac{\partial u_g}{\partial p} \right) = 2 \left(\frac{\partial v_g}{\partial y} \frac{\partial T}{\partial y} + \frac{\partial u_g}{\partial y} \frac{\partial T}{\partial x} \right) = -2 \frac{d_g}{dt} \left(\frac{\partial T}{\partial y} \right) \equiv -2Q_{g2}$$

Error found by Anja: insert minus sign

$$\frac{d_g}{dt} \left(\frac{\partial T}{\partial y} \right) = -\frac{\partial v_g}{\partial y} \frac{\partial T}{\partial y} - \frac{\partial u_g}{\partial y} \frac{\partial T}{\partial x}$$

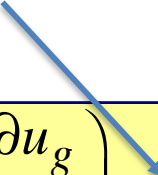
(previous slide)

Dynamical consequences of frontogenesis in a quasi-geostrophic atmosphere

Thermal wind balance:

$$\frac{\partial u_g}{\partial p} = \frac{R}{pf_0} \frac{\partial T}{\partial y} \quad \frac{\partial v_g}{\partial p} = -\frac{R}{pf_0} \frac{\partial T}{\partial x}$$

Error found by Anja: insert minus sign



$$\frac{d_g}{dt} \left(\frac{\partial T}{\partial y} - \frac{f_0 p}{R} \frac{\partial u_g}{\partial p} \right) = 2 \left(\frac{\partial v_g}{\partial y} \frac{\partial T}{\partial y} + \frac{\partial u_g}{\partial y} \frac{\partial T}{\partial x} \right) = -2 \frac{d_g}{dt} \left(\frac{\partial T}{\partial y} \right) \equiv -2Q_{g2}$$

This equation is consistent with thermal wind balance only if $Q_{g2}=0$.

But, remember that we have neglected ageostrophic motion

Now if we include ageostrophic motion and repeat the derivation, we conclude that ageostrophic motion is needed to maintain thermal wind balance !

The role of **a**geostrophic flow is to preserve thermal wind balance

Neglecting ageostrophic flow we have (previous slides) :

$$\frac{d_g}{dt} \left(\frac{f_0 p}{R} \frac{\partial u_g}{\partial p} - \frac{\partial T}{\partial y} \right) = -2Q_{g_2}$$

Repeating the derivation of the previous slides including ageostrophic flow yields

$$\frac{d_g}{dt} \left(\frac{f_0 p}{R} \frac{\partial u_g}{\partial p} - \frac{\partial T}{\partial y} \right) = -2Q_{g_2} + \frac{f_0^2 p}{R} \frac{\partial v_a}{\partial p} - \frac{p S_p}{R} \frac{\partial \omega}{\partial y}$$

($\beta=0$ and $J=0$)

The role of **a**geostrophic flow is to preserve thermal wind balance

Neglecting ageostrophic flow we have (previous slides) :

$$\frac{d_g}{dt} \left(\frac{f_0 p}{R} \frac{\partial u_g}{\partial p} - \frac{\partial T}{\partial y} \right) = -2Q_{g_2}$$

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$$\frac{d_g}{dt} \left(\frac{f_0 p}{R} \frac{\partial u_g}{\partial p} - \frac{\partial T}{\partial y} \right) = -2Q_{g_2} + \frac{f_0^2 p}{R} \frac{\partial v_a}{\partial p} - \frac{p S_p}{R} \frac{\partial \omega}{\partial y} = 0$$

“Conservation” of thermal wind balance

$$-2Q_{g_2} + \frac{f_0^2 p}{R} \frac{\partial v_a}{\partial p} - \frac{p S_p}{R} \frac{\partial \omega}{\partial y} = 0$$

to next slide

The role of **a**geostrophic flow is to preserve thermal wind balance

Previous slide:

$$-2Q_{g_2} + \frac{f_0^2 p}{R} \frac{\partial v_a}{\partial p} - \frac{p S_p}{R} \frac{\partial \omega}{\partial y} = 0$$

eq.1.355

Likewise, the x-component of thermal wind balance yields:

$$-2Q_{g_1} + \frac{f_0^2 p}{R} \frac{\partial u_a}{\partial p} - \frac{p S_p}{R} \frac{\partial \omega}{\partial x} = 0$$

eq.1.356

The role of **a**geostrophic flow is to preserve thermal wind balance

$$\frac{\partial}{\partial y} \left[-2Q_{g_2} + \frac{f_0^2 p}{R} \frac{\partial v_a}{\partial p} - \frac{p S_p}{R} \frac{\partial \omega}{\partial y} \right] = 0$$

+

$$\frac{\partial}{\partial x} \left[-2Q_{g_1} + \frac{f_0^2 p}{R} \frac{\partial u_a}{\partial p} - \frac{p S_p}{R} \frac{\partial \omega}{\partial x} \right] = 0$$



From the two equations above:

$$S_p \nabla_h^2 \omega + f_0^2 \frac{\partial^2 \omega}{\partial p^2} = -\frac{2R}{p} \vec{\nabla} \cdot \vec{Q}_g$$

eq. 1.359

("omega equation")

we have used

$$\frac{\partial u_a}{\partial x} + \frac{\partial v_a}{\partial y} + \frac{\partial \omega}{\partial p} = 0$$

and

$$\nabla_h^2 \equiv \frac{\partial^2}{\partial x^2} + \frac{\partial^2}{\partial y^2}$$

The omega equation

$$S_p \nabla_h^2 \omega + f_0^2 \frac{\partial^2 \omega}{\partial p^2} = -\frac{2R}{p} \vec{\nabla} \cdot \vec{Q}_g \quad \text{eq. 1.359}$$

x-component Q-vector:

y-component Q-vector:

$$Q_{g1} = -\left(\frac{\partial u_g}{\partial x} \frac{\partial T}{\partial x} + \frac{\partial v_g}{\partial x} \frac{\partial T}{\partial y} \right); \quad Q_{g2} = -\left(\frac{\partial u_g}{\partial y} \frac{\partial T}{\partial x} + \frac{\partial v_g}{\partial y} \frac{\partial T}{\partial y} \right).$$

Since both T , u_g and v_g can all be expressed as a function of Φ , we can calculate the vertical motion from the distribution of Φ only!!!

With the hydrostatic approximation and the equation of state, show that T can be expressed as a function of Φ .

The omega equation: interpretation

$$S_p \nabla_h^2 \omega + f_0^2 \frac{\partial^2 \omega}{\partial p^2} = -\frac{2R}{p} \vec{\nabla} \cdot \vec{Q}_g \longrightarrow$$

eq. 1.359

The omega equation is a Poisson-like equation, i.e. second order elliptic partial differential equation.

The left hand side is proportional to $-\omega$ (see eq. 1.363).

Therefore, a simplified version of the omega equation is

$$\longrightarrow -\omega \approx w \approx -\vec{\nabla} \cdot \vec{Q}_g \quad \text{eq. 1.363}$$

i.e. **upward** (**downward**) motion if Q_g -vector is **convergent** (**divergent**)

Illustrated on the next slide 

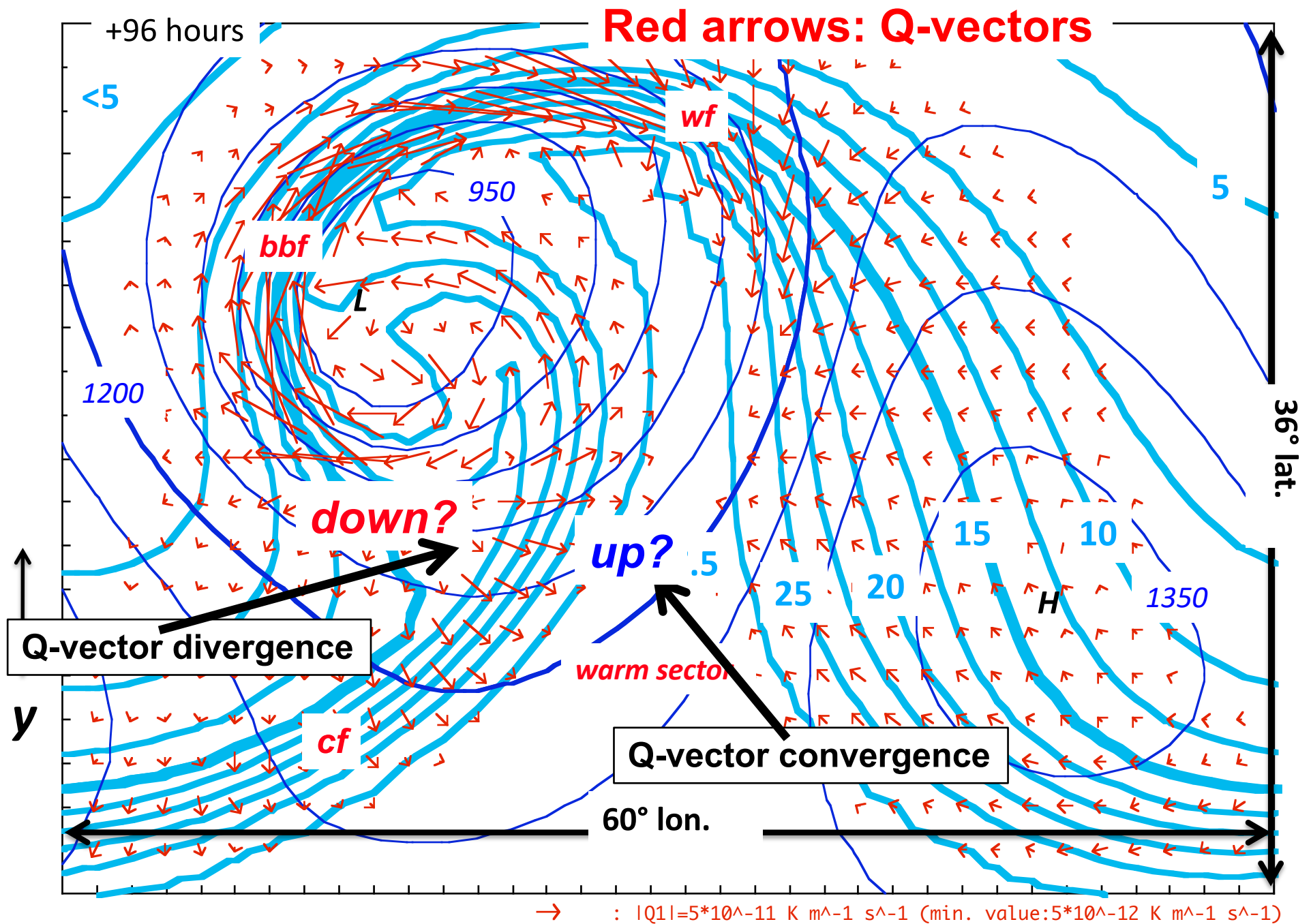
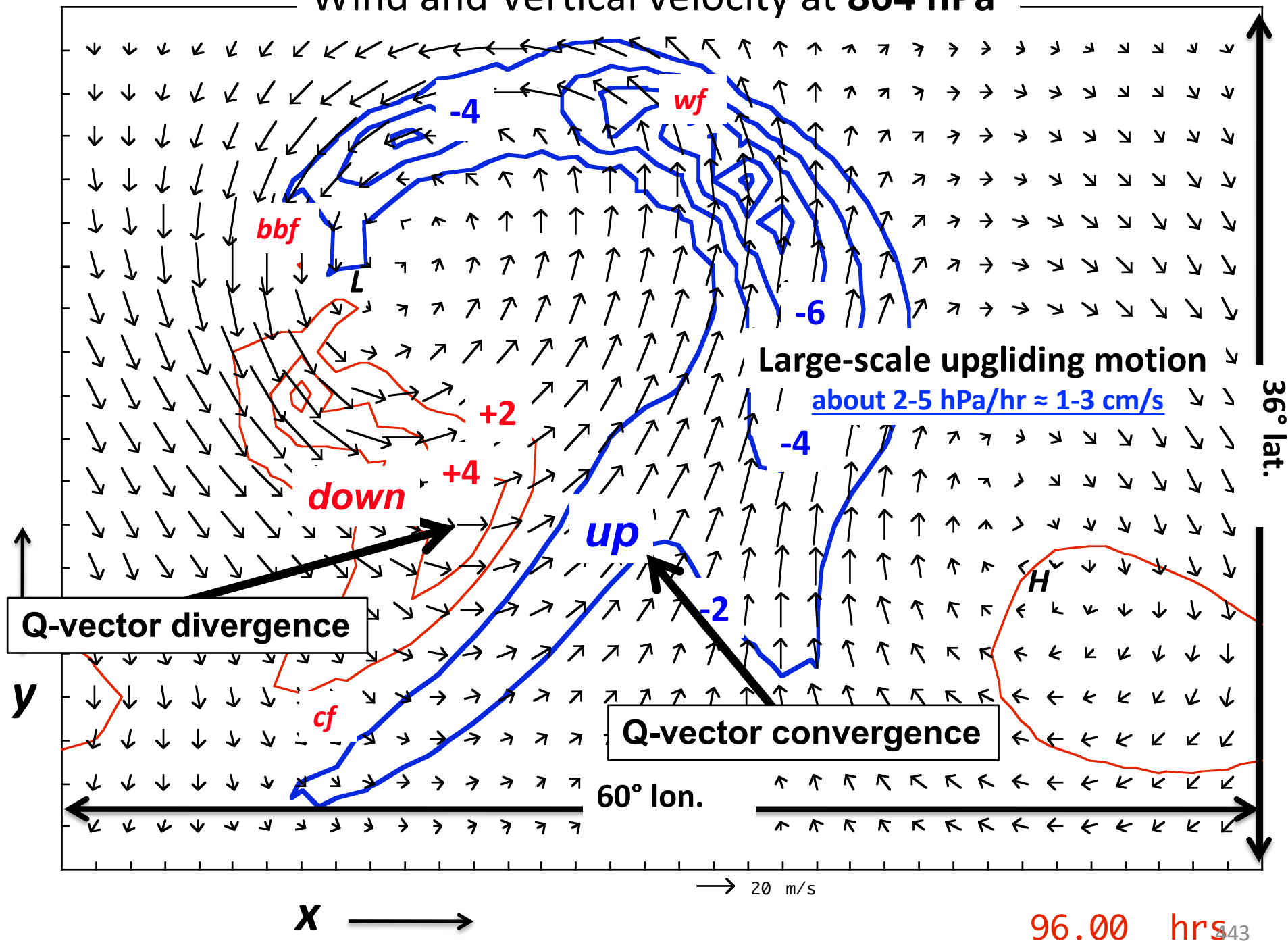
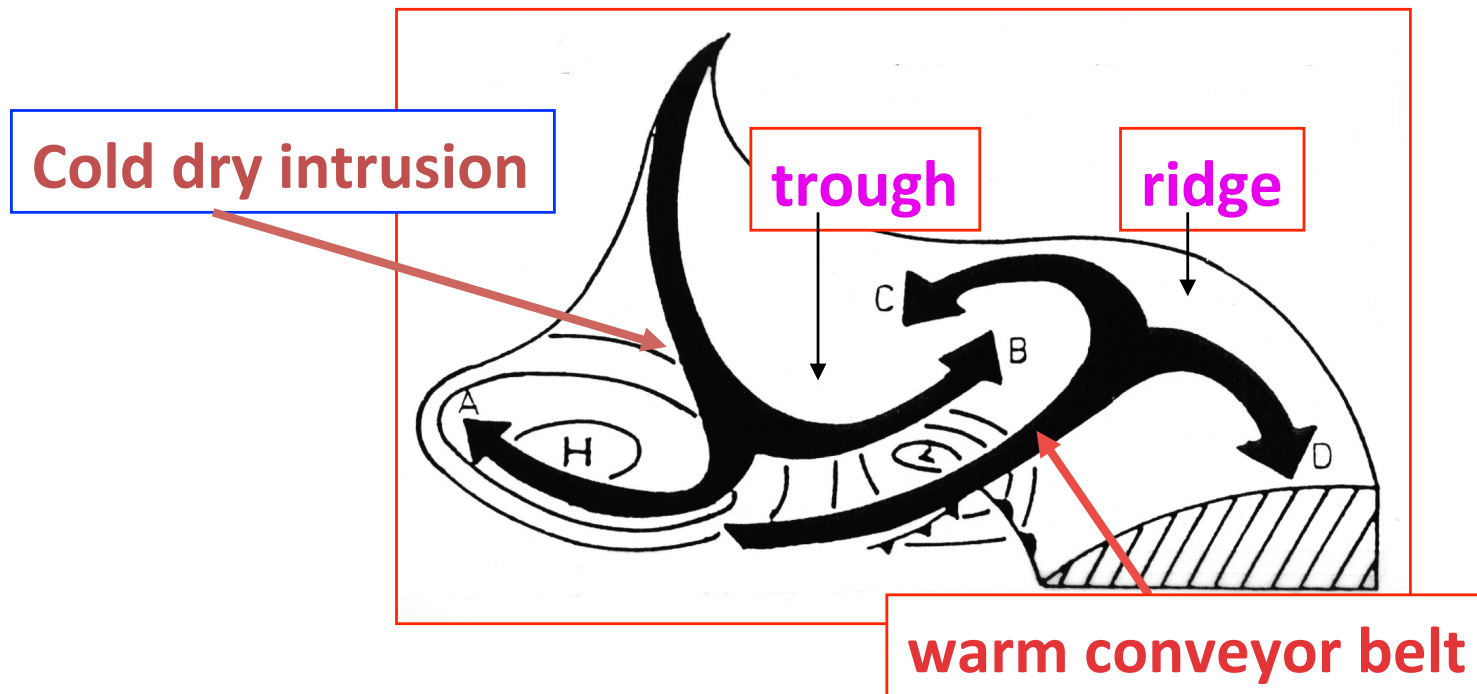


Figure 1.128

Wind and Vertical velocity at 864 hPa



Conceptual model



Schematic of relative flow within a baroclinic wave. Solid arrows represent flow along a sloping isentropic surface. The surface pressure pattern is indicated. Flow is very *slanted*. This can be explained also as a consequence of the preferred large value of aspect ratio.

(Thorncroft, C.D., B.J. Hoskins and M.E. McIntyre, 1993: Two paradigms of baroclinic-wave life-cycle behaviour. **Q.J.R.Meteorol.Soc.**, 119, 17-55).

Apply the omega equation to a hypothetical case

The pattern of vertical motion

- According to the omega equation (relation to frontogenesis)
- Relation to PV-inversion theory: the “vacuum-cleaner effect”

Omega equation:

$$S_p \nabla_h^2 \omega + f_0^2 \frac{\partial^2 \omega}{\partial p^2} = -\frac{2R}{p} \vec{\nabla} \cdot \vec{Q}_g \quad \text{eq. 1.359}$$

Static stability

x-component Q-vector:

y-component Q-vector:

$$Q_{g1} = -\left(\frac{\partial u_g}{\partial x} \frac{\partial T}{\partial x} + \frac{\partial v_g}{\partial x} \frac{\partial T}{\partial y} \right); \quad Q_{g2} = -\left(\frac{\partial u_g}{\partial y} \frac{\partial T}{\partial x} + \frac{\partial v_g}{\partial y} \frac{\partial T}{\partial y} \right).$$

(lecture 15)

Determine the two components of the Q-vector for the distribution of the geopotential given in **problem 1.35**

Problem 1.35: theoretical case

The **geopotential field** is hypothetically given by

$$\Phi = \Phi_0(p) + Uf_0 \left\{ -y \left[\cos\left(\frac{\pi p}{p_{ref}}\right) + 1 \right] + \frac{1}{l} \sin(l(x - ct)) \right\} \quad (\text{linear decrease with } y)$$

In Q-G-theory geopotential, Φ , determines everything else...

The **geostrophic wind**:

$$u_g = -\frac{1}{f_0} \frac{\partial \Phi}{\partial y} = U \left\{ \cos\left(\frac{\pi p}{p_{ref}}\right) + 1 \right\} \rightarrow \frac{\partial u_g}{\partial x} = 0 \quad ; \quad \frac{\partial u_g}{\partial y} = 0$$

$$v_g = \frac{1}{f_0} \frac{\partial \Phi}{\partial x} = U \cos(l(x - ct)) \rightarrow \frac{\partial v_g}{\partial x} = -Ul \sin(l(x - ct)) \quad ; \quad \frac{\partial v_g}{\partial y} = 0$$

$$Q_{g1} = -\left(\frac{\partial u_g}{\partial x} \frac{\partial T}{\partial x} + \frac{\partial v_g}{\partial x} \frac{\partial T}{\partial y} \right) = -\frac{\partial v_g}{\partial x} \frac{\partial T}{\partial y}$$

$$Q_{g2} = -\left(\frac{\partial u_g}{\partial y} \frac{\partial T}{\partial x} + \frac{\partial v_g}{\partial y} \frac{\partial T}{\partial y} \right) = 0$$

Meridional temperature gradient

Zonal **geostrophic wind**:

$$u_g = -\frac{1}{f_0} \frac{\partial \Phi}{\partial y} = U \left\{ \cos\left(\frac{\pi p}{p_{ref}}\right) + 1 \right\} \quad (\text{previous slide})$$

(u_g depends on p)

Temperature gradient is determined by **thermal wind relation** (1.347):

$$\frac{\partial u_g}{\partial p} = \frac{R}{pf_0} \frac{\partial T}{\partial y} \quad \longrightarrow \quad \frac{\partial T}{\partial y} = -\frac{\pi U f_0}{R} \frac{p}{p_{ref}} \sin\left(\frac{\pi p}{p_{ref}}\right) < 0$$

Previous slide:

$$\frac{\partial v_g}{\partial x} = \zeta_g = -Ul \sin(l(x - ct))$$

$$\longrightarrow Q_{g1} = -\zeta_g \frac{\partial T}{\partial y} = -Ul \sin(l(x - ct)) \frac{\pi U f_0}{R} \frac{p}{p_{ref}} \sin\left(\frac{\pi p}{p_{ref}}\right) \quad Q_{g2} = 0$$

forcing associated with rotation of the isotherms by the flow

Omega equation

Previous slide:

$$Q_{g1} = -\frac{\pi f_0 U^2 l}{R} \frac{p}{p_{ref}} \sin\left(\frac{\pi p}{p_{ref}}\right) \sin(l(x - ct))$$

$$Q_{g2} = 0$$

Omega equation (lecture 15):

$$S_p \nabla_h^2 \omega + f_0^2 \frac{\partial^2 \omega}{\partial p^2} = -\frac{2R}{p} \vec{\nabla} \cdot \vec{Q}_g$$

The 2D (x,p) -approximation of the omega equation is, after substitution of the expression for Q_{g1}

$$\frac{\partial^2 \omega}{\partial x^2} + \frac{f_0^2}{S_p} \frac{\partial^2 \omega}{\partial p^2} = \frac{2\pi f_0 U^2 l^2}{S_p p_{ref}} \sin\left(\frac{\pi p}{p_{ref}}\right) \cos(l(x - ct))$$

Left hand side is proportional to $-\omega$ or $+\omega$ *. Hence we obtain the following approximate expression:

$$\omega \propto \frac{2\pi f_0 U^2 l^2}{S_p p_{ref}} \sin\left(\frac{\pi p}{p_{ref}}\right) \cos(l(x - ct))$$

* p. 263

Vertical motion according to the omega equation

Previous slide:

$$Q_{g1} = -\zeta_g \frac{\partial T}{\partial y} = -\frac{\pi f_0 U^2 l}{R} \frac{p}{p_{ref}} \sin\left(\frac{\pi p}{p_{ref}}\right) \sin(l(x - ct))$$

$$Q_{g2} = 0$$

$$w \propto \frac{2\pi f_0 U^2 l^2}{S_p p_{ref}} \sin\left(\frac{\pi p}{p_{ref}}\right) \cos(l(x - ct))$$

Remember: $v_g = U \cos(l(x - ct))$
 $\zeta_g = -Ul \sin(l(x - ct))$

- w is in phase with v_g :

poleward motion is upward and equatorward motion is downward

- w is proportional to l^2 (the square of the wave number)

vertical motion is stronger in short waves than in long waves

- w is out of phase with relative vorticity

Vertical motion is zero in the trough and in the ridge

The Rossby wave case of July 1996

Lecture 11:

Rossby waves in July 1996

Fig. 1.111

Zonal wavelength $\approx 36^\circ$
(wave number, $k \approx 10$)

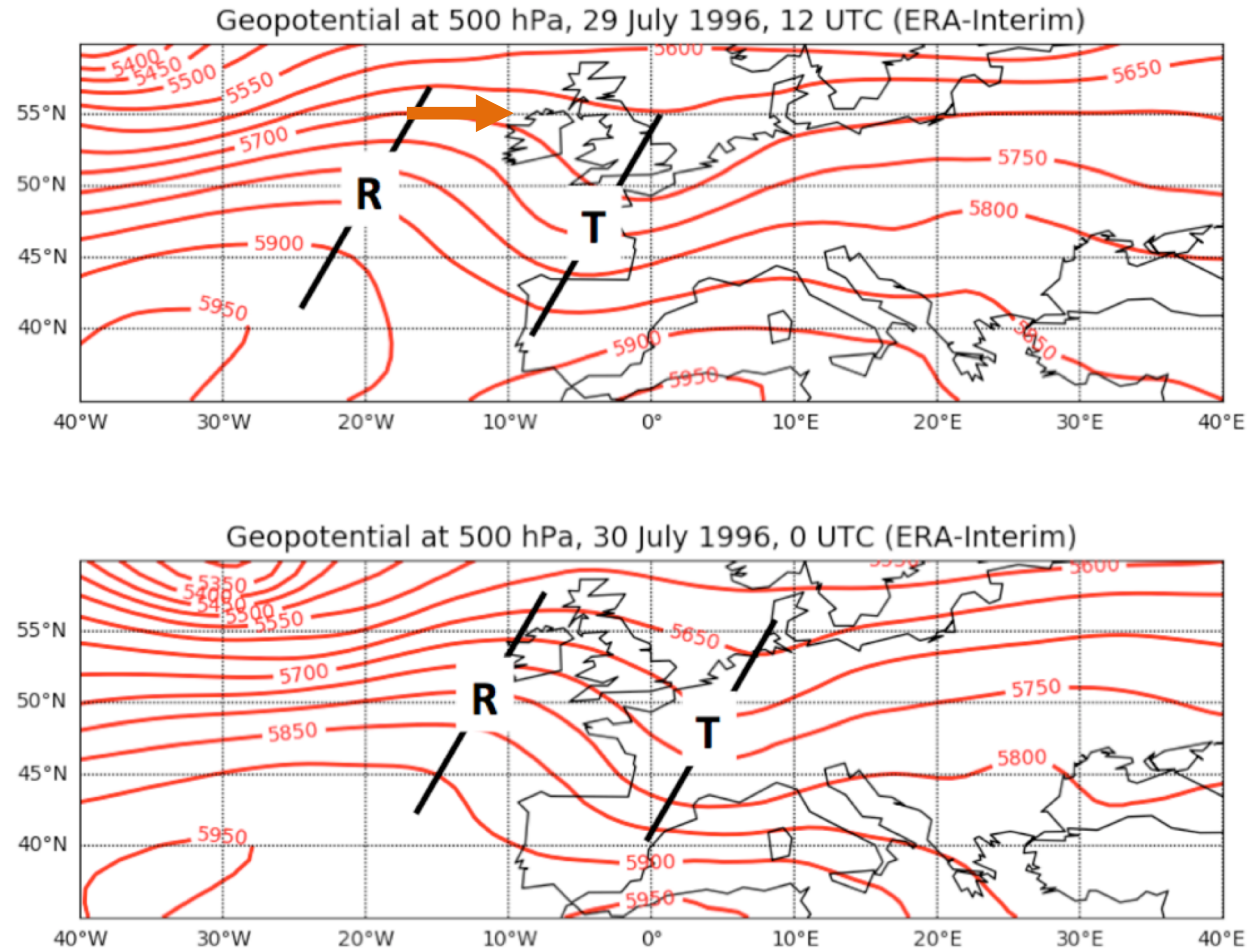
Phase velocity, $c = 10.5 \text{ m s}^{-1}$

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

$$c = \frac{\omega}{l}$$

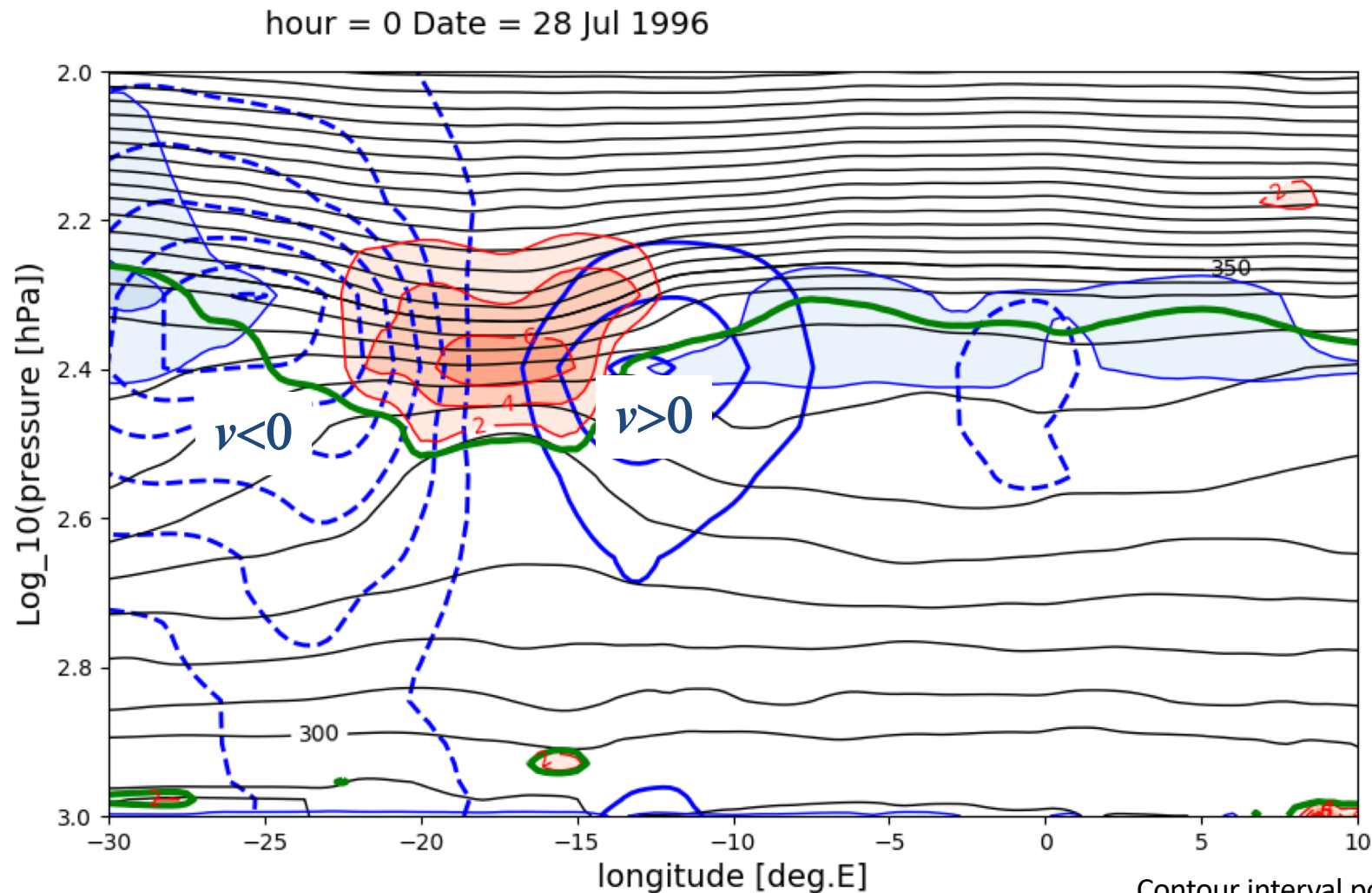
$$l = \frac{2\pi k}{L}$$

Circumference of the earth at certain latitude (ϕ)



The planetary wave of July 1996: θ , v and PV-anomaly

Labels in PVU



(see lecture 16)

Contour interval pot. temp.: 5 K

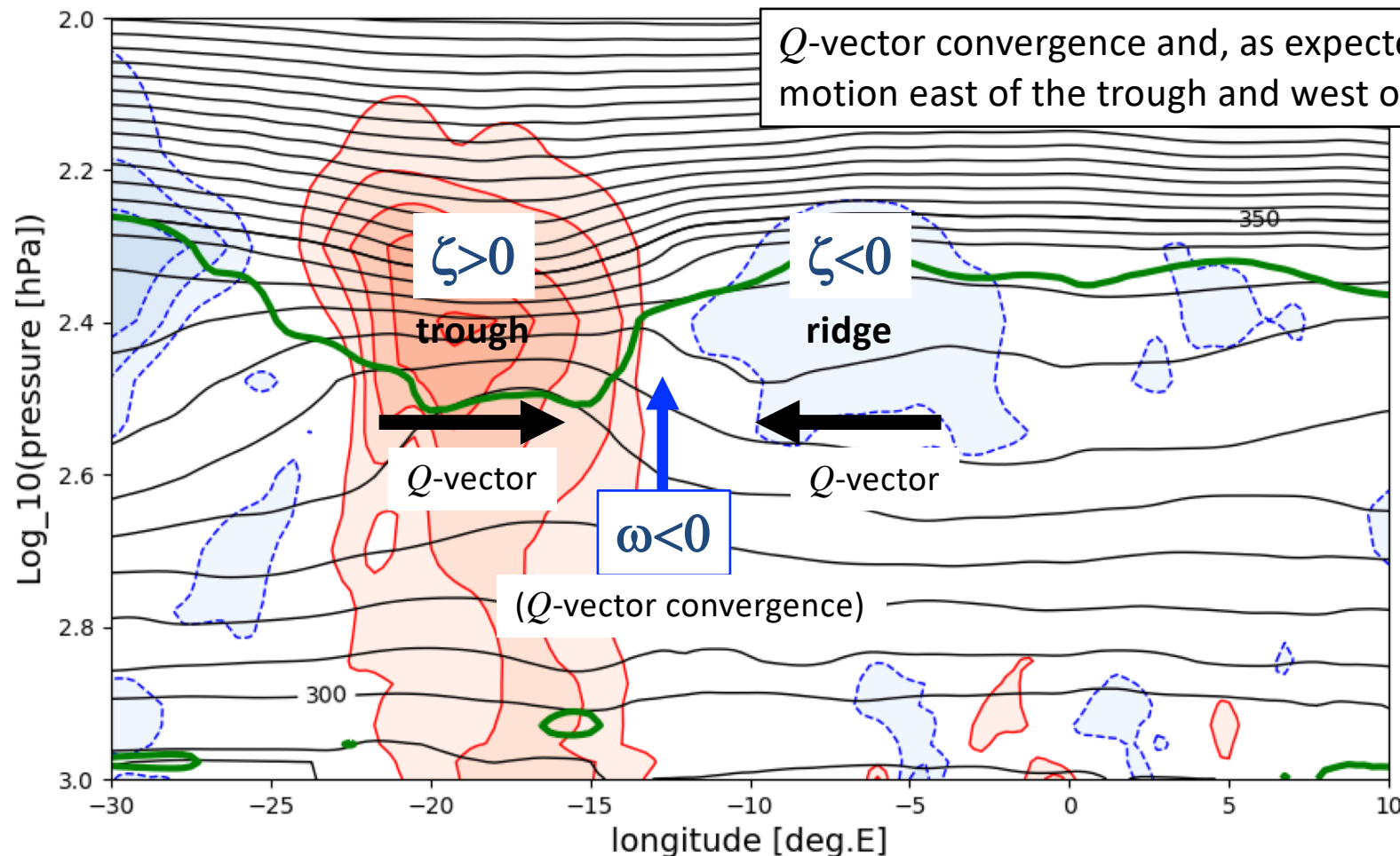
Contour interval v : 5 m s⁻¹

451

The planetary wave of July 1996: θ and ζ

hour = 0 Date = 28 Jul 1996

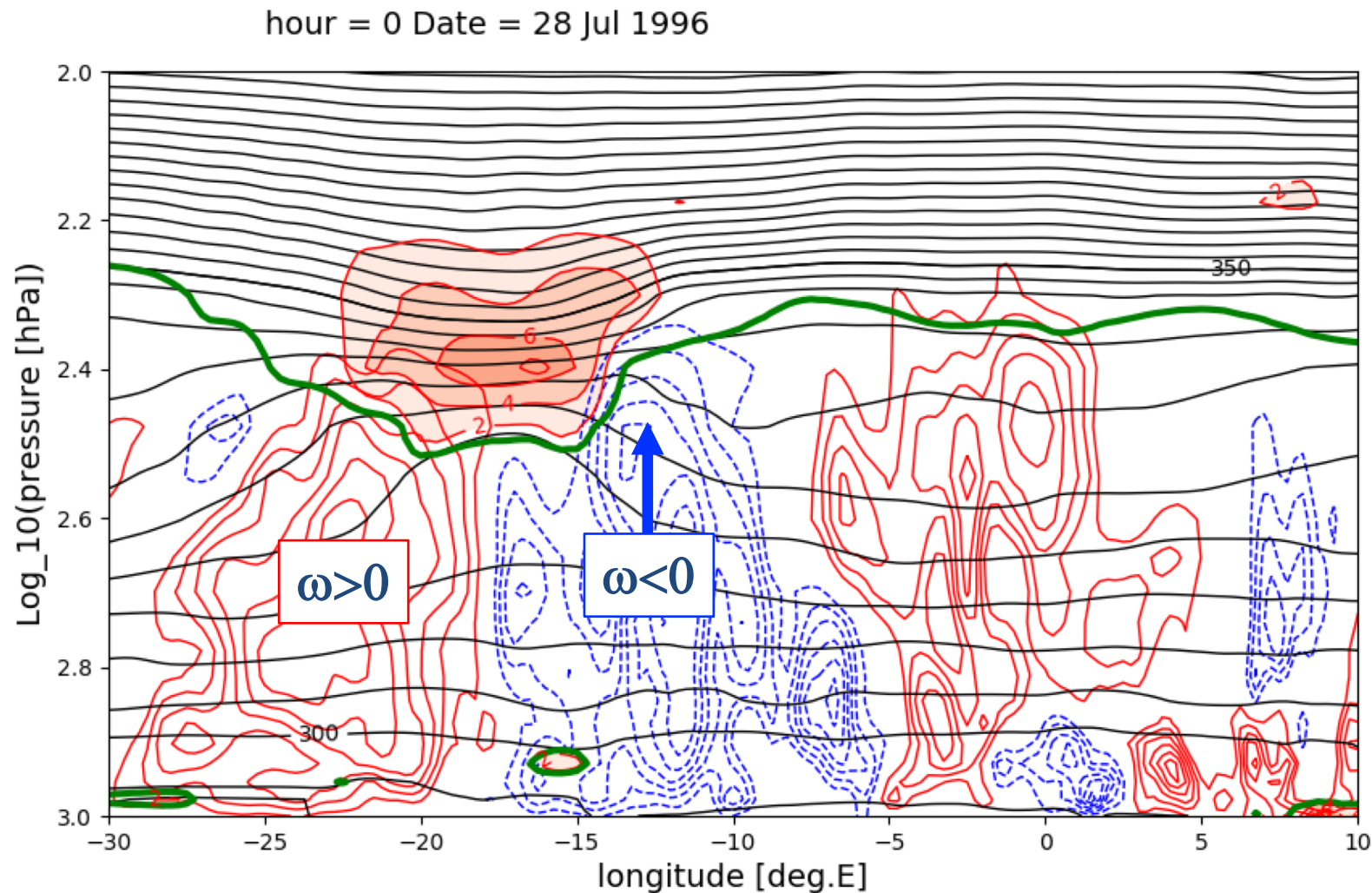
$$Q_{g1} = -\zeta_g \frac{\partial T}{\partial y} \begin{cases} \text{Q-vector points eastwards in the trough} \\ \text{Q-vector points westwards in the ridge} \end{cases}$$



(see lecture 16)

Contour interval vorticity: $0.2 \times 10^{-4} \text{ s}^{-1}$

The planetary wave of July 1996: θ , ω and PV-anomaly



(see lecture 16)

Contour interval ω : 0.02 Pa s⁻¹

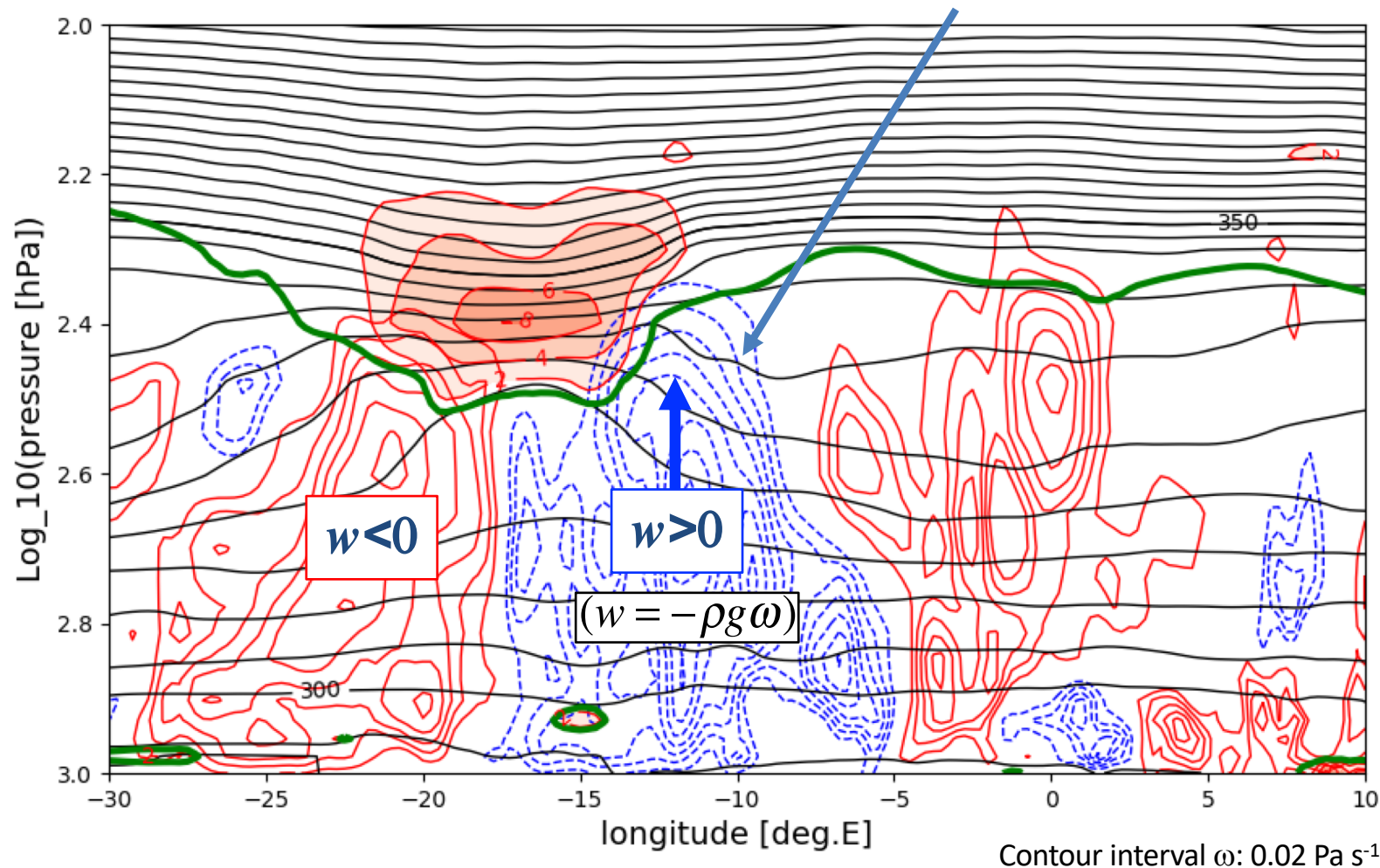
The planetary wave of July 1996: θ , ω and PV-anomaly

Note: upward bulging isentropes below the PV-anomaly!:

In advance of the propagating PV-anomaly we see that:

hour = 1 Date = 28 Jul 1996

$$\frac{\partial z_{\theta}}{\partial t} > 0$$

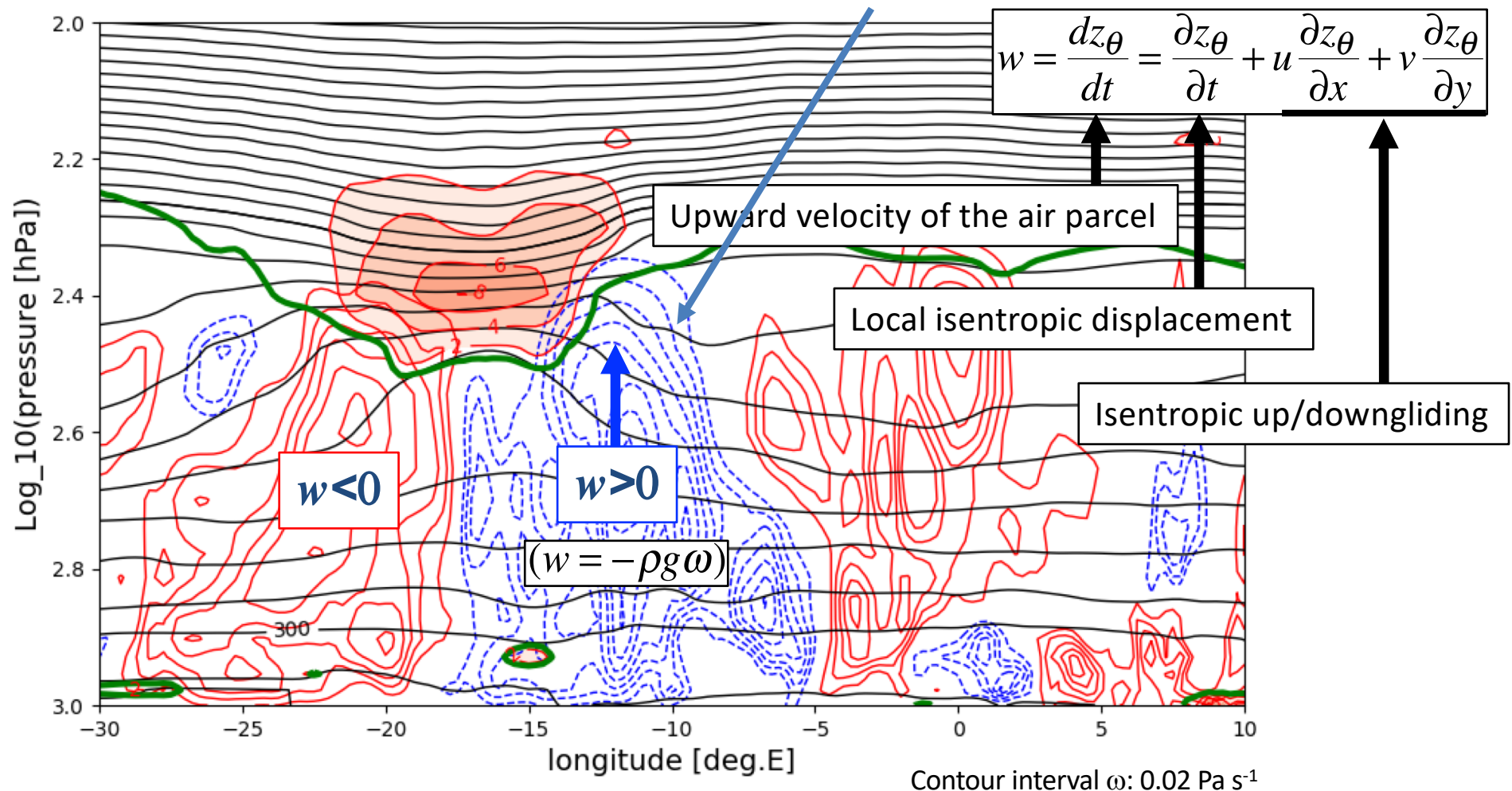


(see lecture 16)

The planetary wave of July 1996: θ , ω and PV-anomaly

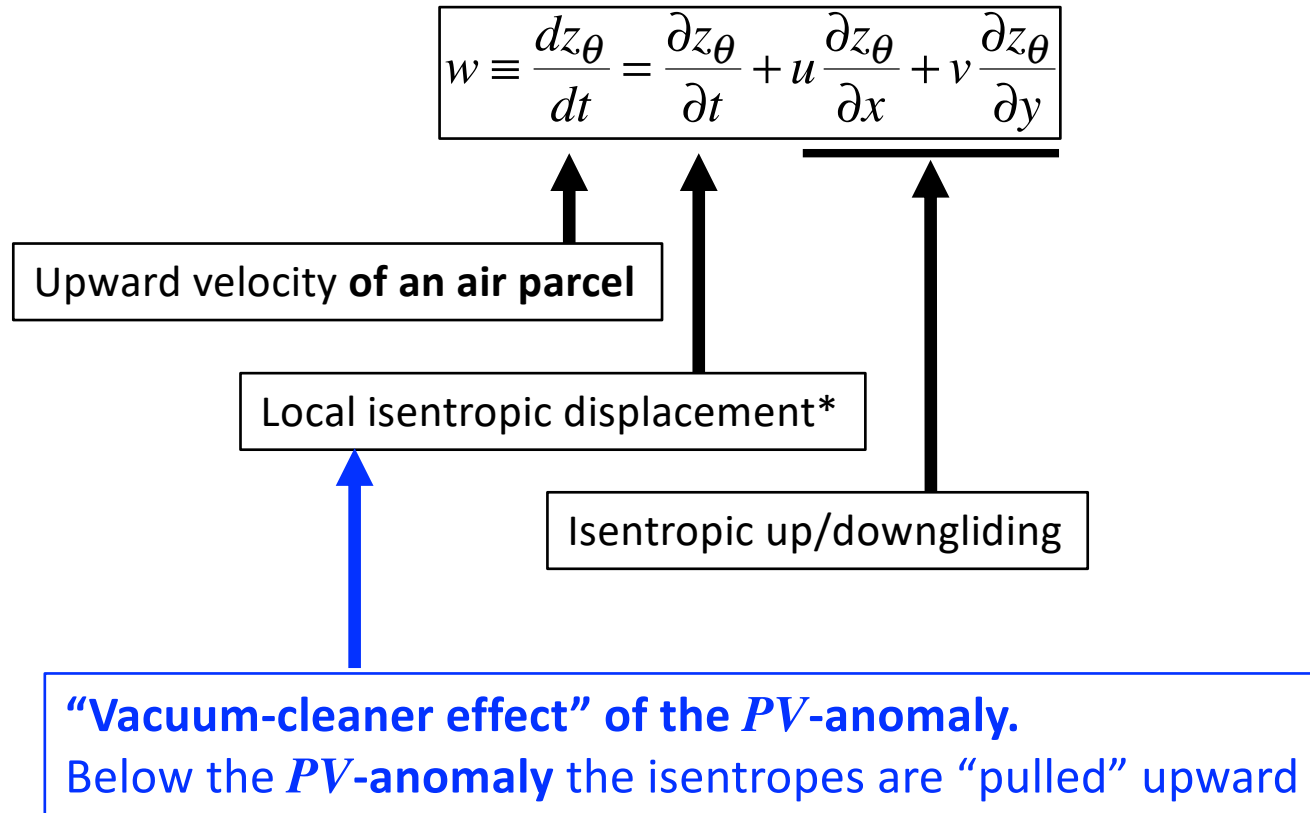
Note: upward bulging isentropes below the PV-anomaly!:

hour = 1 Date = 28 Jul 1996 In advance of the propagating PV-anomaly: $\frac{\partial z_\theta}{\partial t} > 0$



(see lecture 16)

Components of vertical motion: “vacuum cleaner effect”

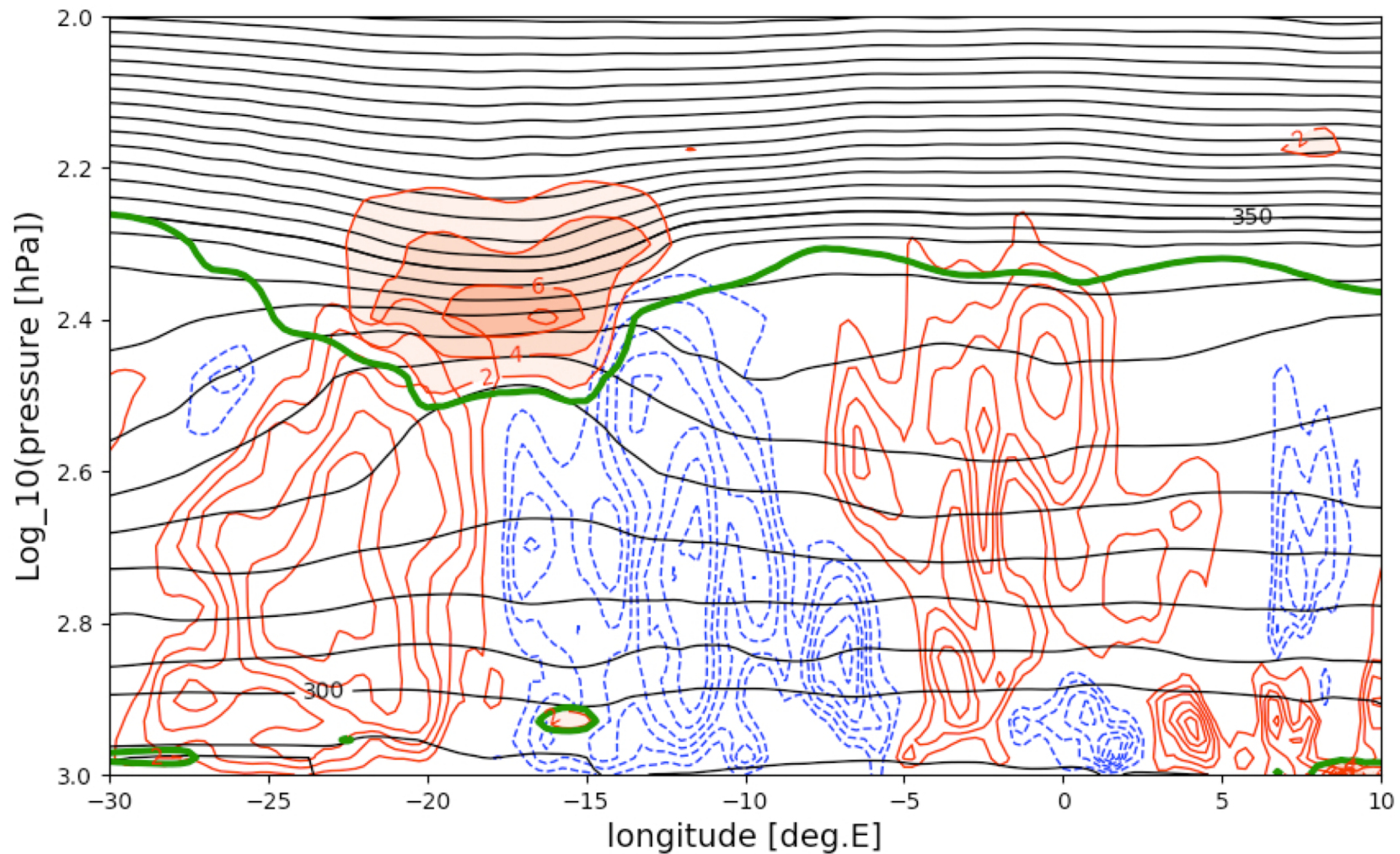


*this is what we see in an animation (next slide)

Animation: watch the isentropes (black contours)

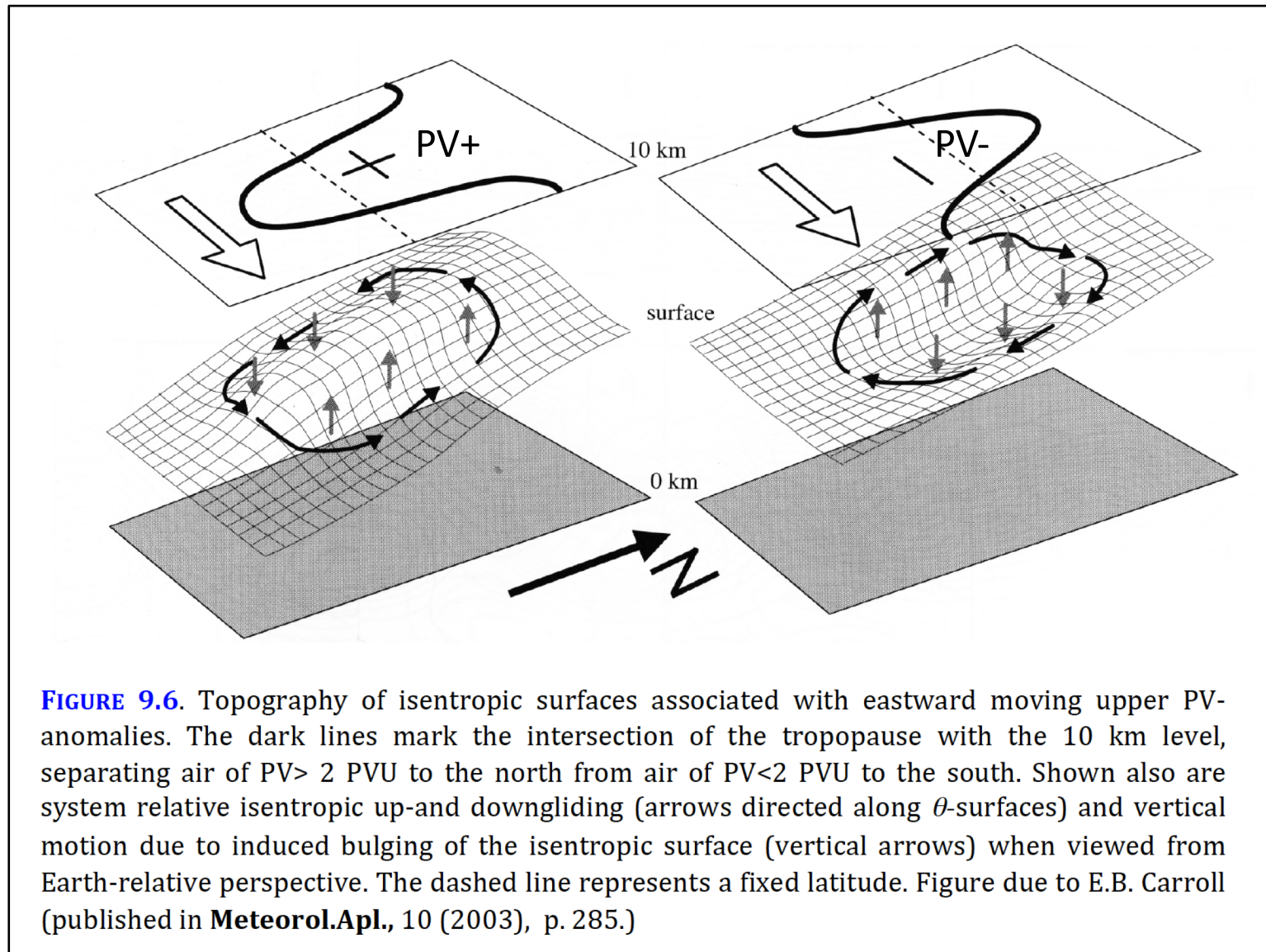
wJuly96_LonCS.mov

hour = 0 Date = 28 Jul 1996



Contour interval ω : 0.02 Pa s⁻¹

“Vacuum-cleaner effect”



Next:

Nearly the End

26/1/2022 (Wednesday, 13:15-15:00)

Overview of the content of the course

Weather (forecast) of 23 January 2022

Forecasting sunshine by far most difficult !

Awards for the winner of the weather forecasting competition

Group photograph

Exam: 2/1/2022 (Wednesday - afternoon)

(a.j.vandelden@uu.nl)

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