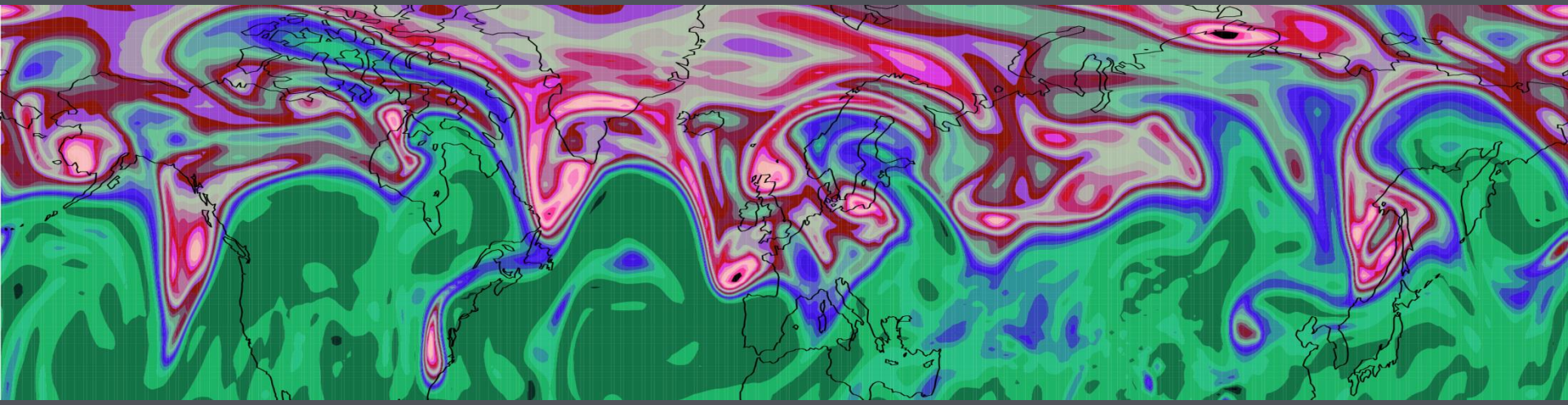
Department of Meteorology

University of Reading

Mechanisms for Diabatic Influence on the Jet Stream



John Methven, Ben Harvey, Leo Saffin, Jake Bland,

Sue Gray & Maarten Ambaum

Department of Meteorology, University of Reading

Andreas Schäfler, DLR

Claudio Sanchez, Met Office

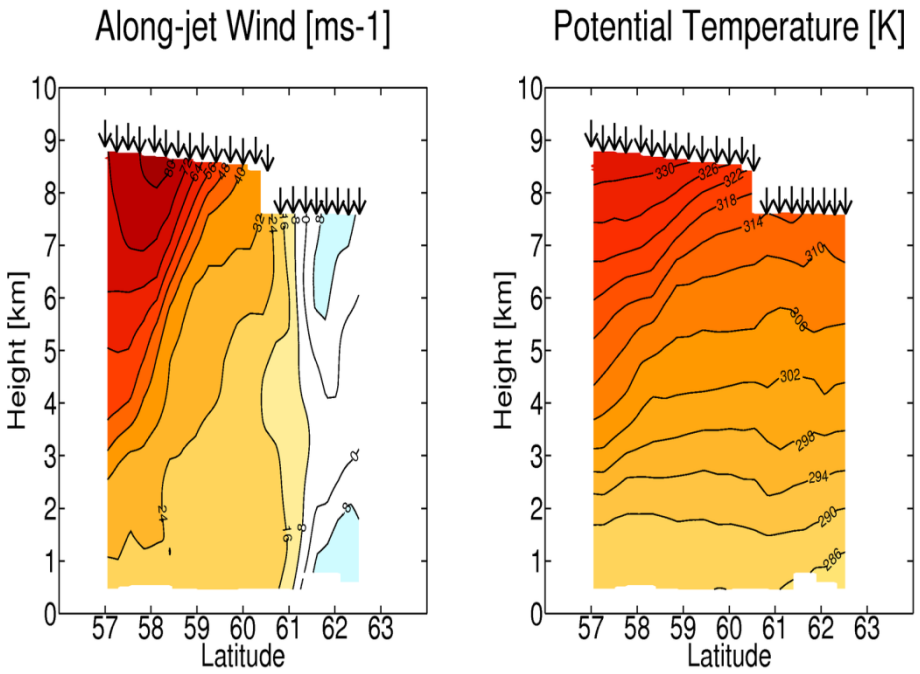
Copyright University of Reading

How can you tell when a model is right – or wrong?

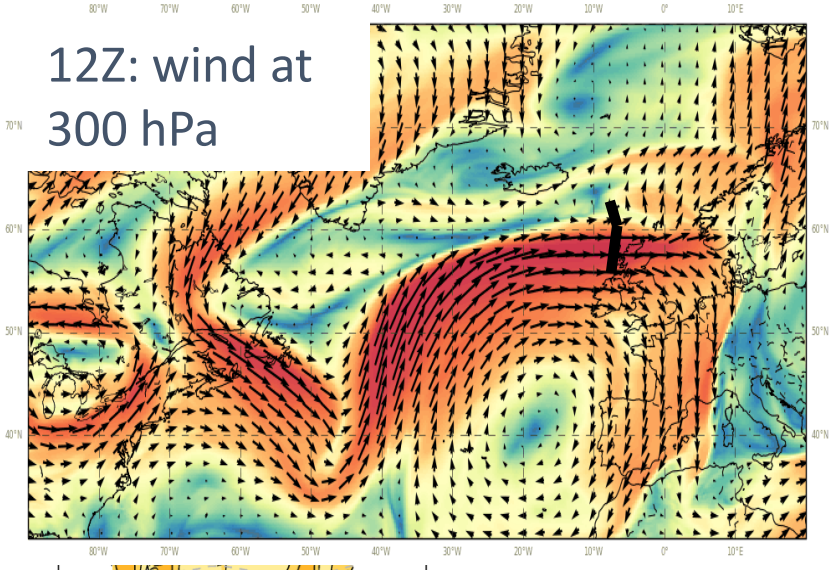
1. **Use physical understanding** to distinguish "physical" and "unphysical" behaviour.
   - Example, potential vorticity (PV) is materially conserved following adiabatic, frictionless flow
   - Furthermore, PV cannot change sign as a result of diabatic mass transport
   - *So, is negative PV (in N. Hemisphere) a sign of model error?*

2. **Use comparison of models (analyses and forecasts) with independent observations**, in many cases, to identify structure of systematic errors, and their ramifications.

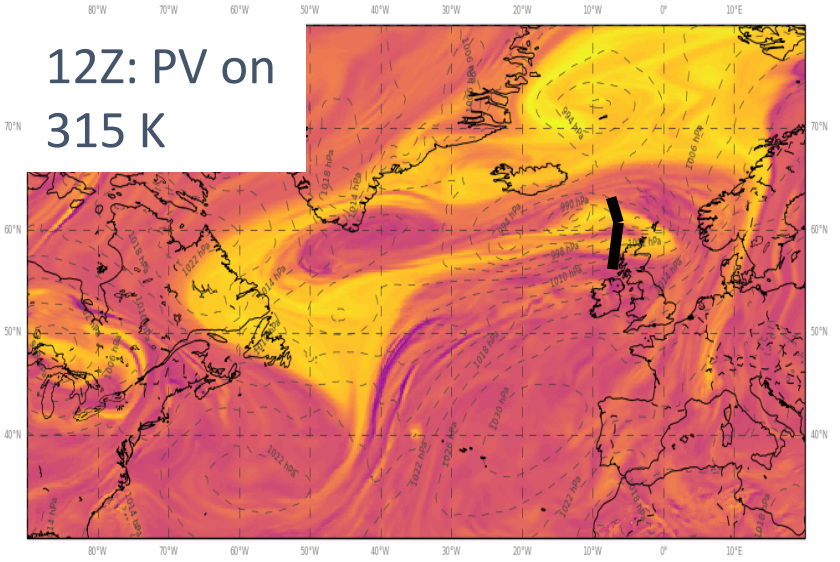
Confronted with new observations: NAWDEX jet streak case (ex-Karl)



12Z: wind at 300 hPa



12Z: PV on 315 K

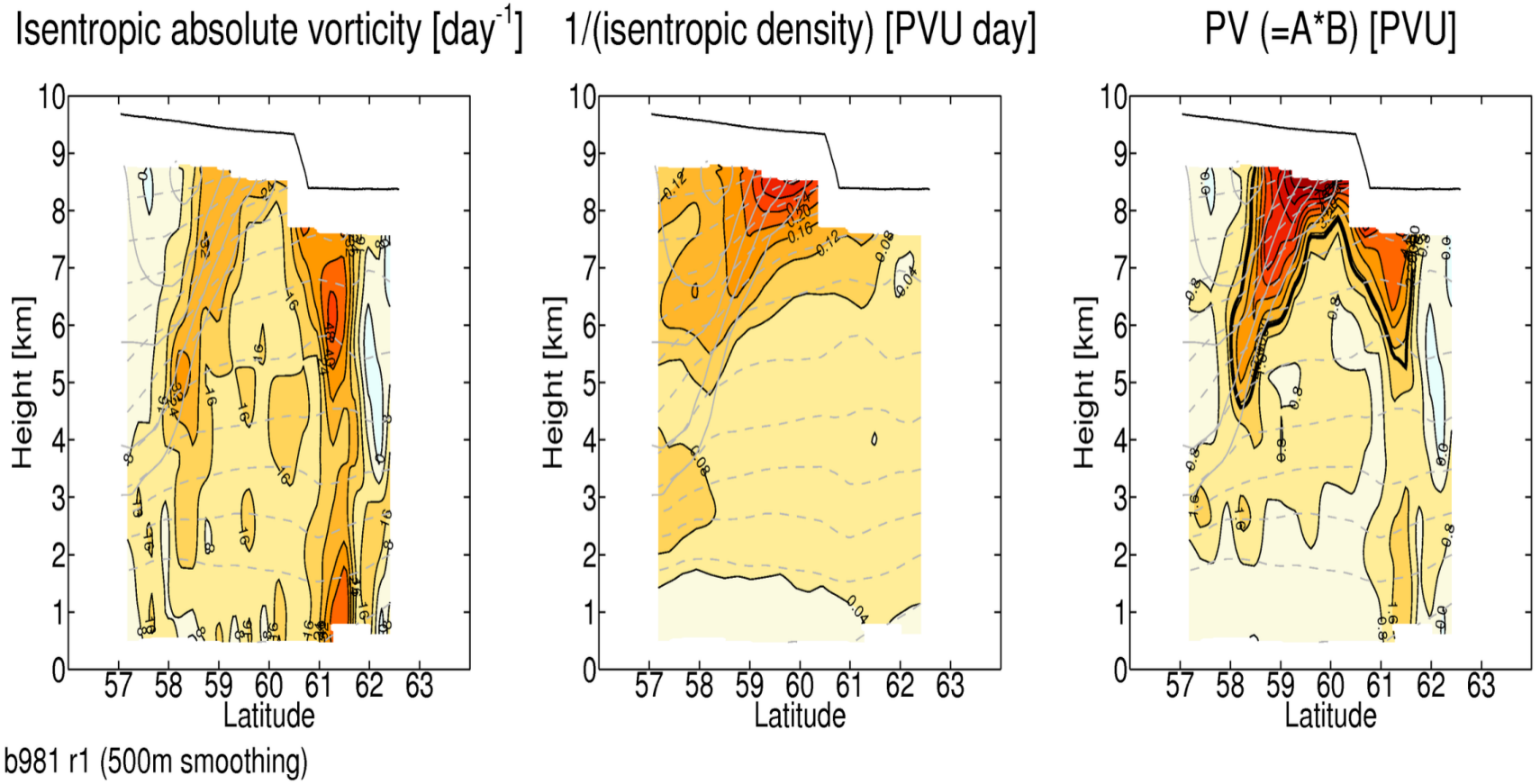


22 dropsondes on section crossing the jet stream

See Ben Harvey's poster

Negative PV observed with dropsondes

Signature of diabatic processes sharpening jet stream



$$P = \frac{1}{\rho}\boldsymbol{\zeta}\cdot\nabla\theta = (\boldsymbol{\zeta}\cdot\boldsymbol{n})\frac{|\nabla\theta|}{\rho} \qquad \text{where} \qquad \boldsymbol{n} = \frac{\nabla\theta}{|\nabla\theta|}$$

Tracking non-conservative changes in PV

University of Reading

- PV tracer diagnostics are based on a Lagrangian decomposition of the PV field.
- Full PV conservation equation can be written:

$$\frac{Dq}{Dt} = \sum_i S_i$$

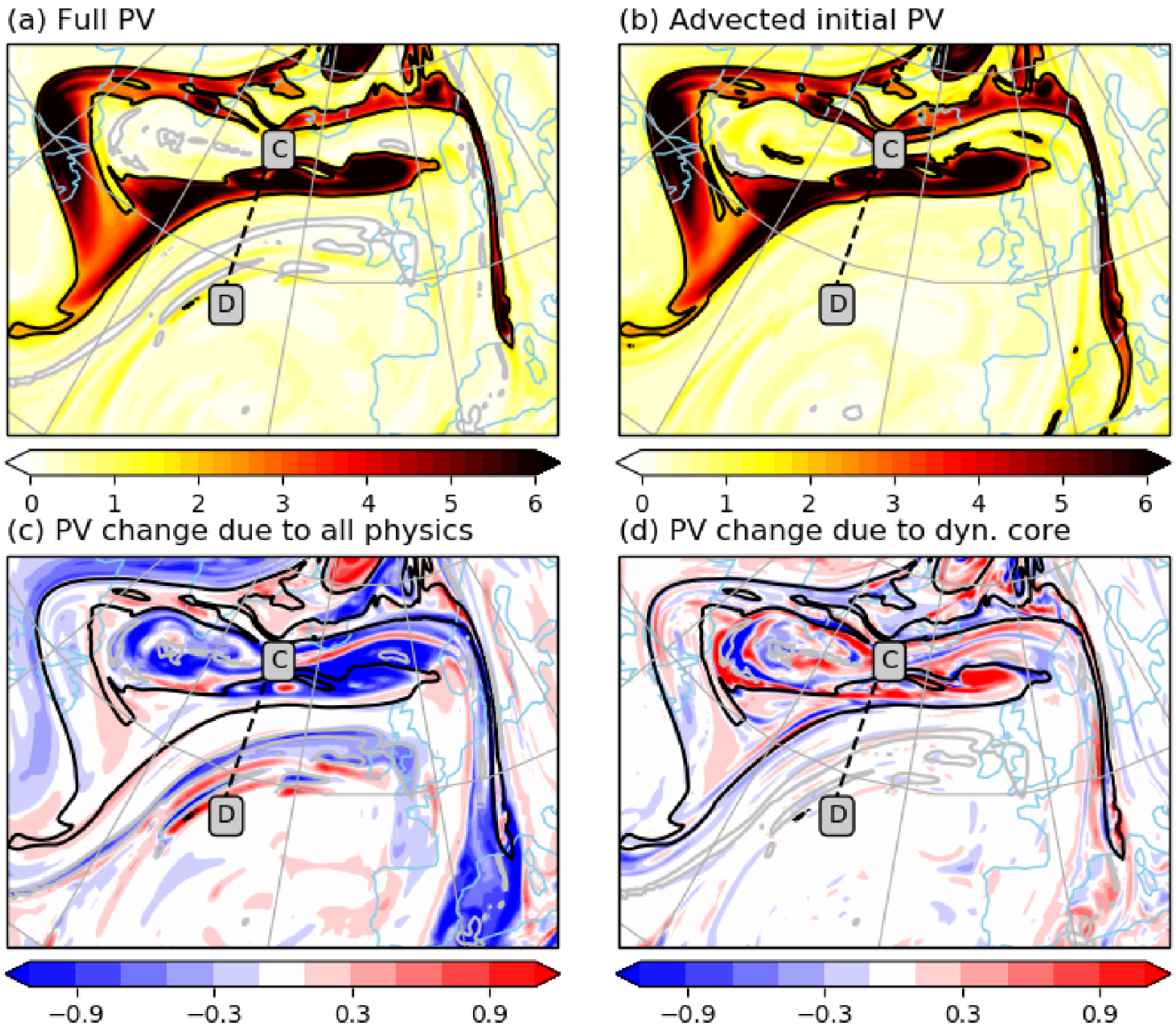
where S_i denotes the Lagrangian tendency resulting from one physical process in model

- **Diabatic tracers** q_i are initialised as zero but experience only one of the S_i terms, and are advected by the semi-Lagrangian scheme.

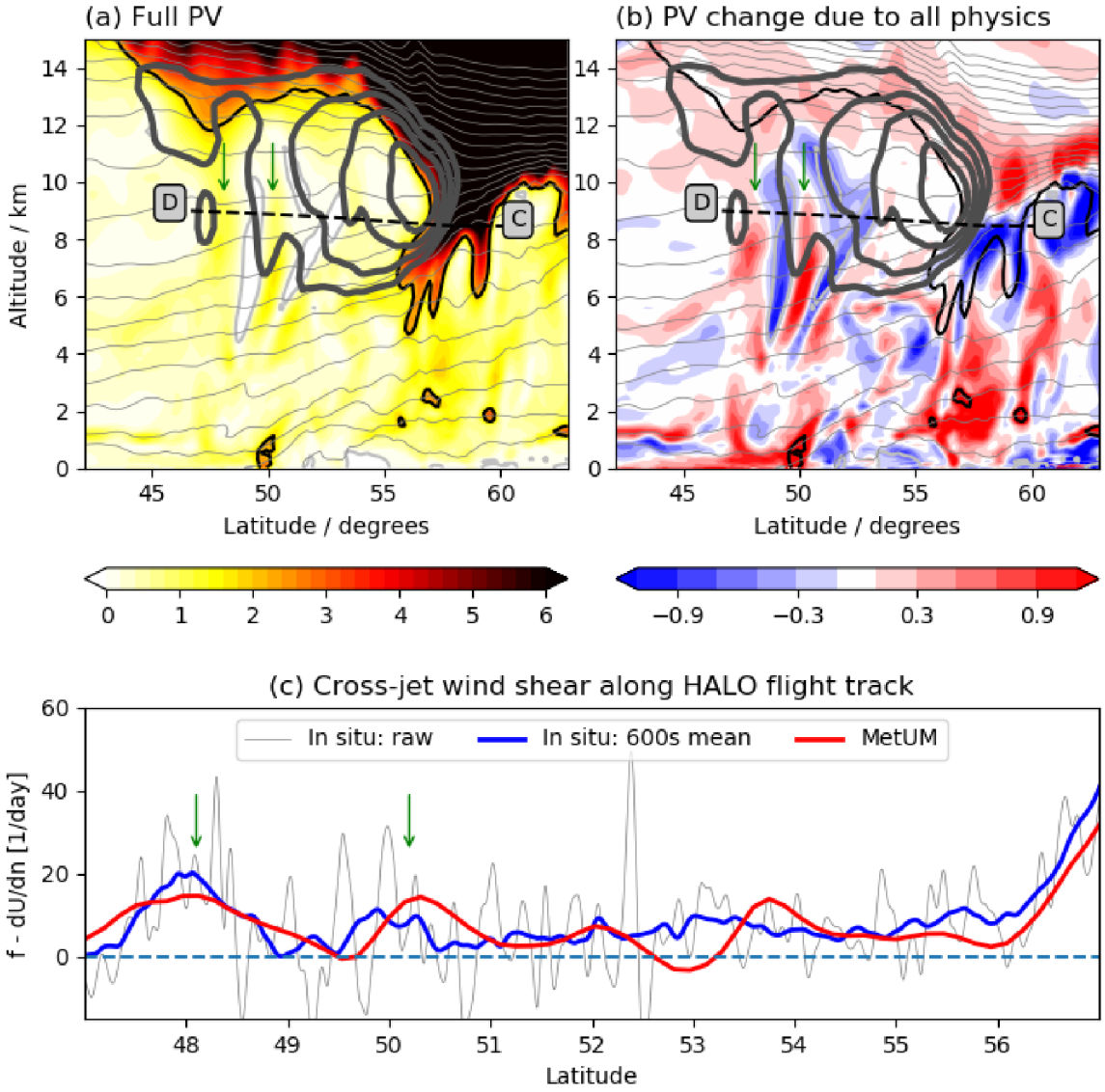
  $\Rightarrow$ each tracer shows accumulated contribution of one process to PV.

$$q = q_{passive} + \sum_i q_i + error \qquad \begin{cases} q_i(t_0) = 0 \\ q_{passive}(t_0) = q(t_0) \end{cases}$$

PV along a section through jet stream and WCB

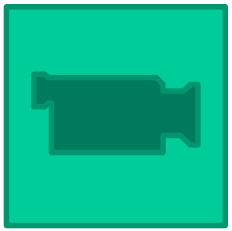


PV along a section through jet stream and WCB



Back to Theory: Jet Stream and Rossby Waves

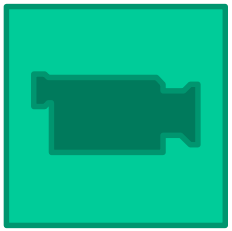
Focus on jet streams at tropopause level



- westerly on average around the extratropics
- highly meandering and variable
- often "split jet regions" with more than one

Why is there so much wave activity on the jet stream?

- jet maximum = strong **potential vorticity (PV) gradient**
- Rossby waves propagate on PV gradients
- Shear instability grows through Rossby wave interaction



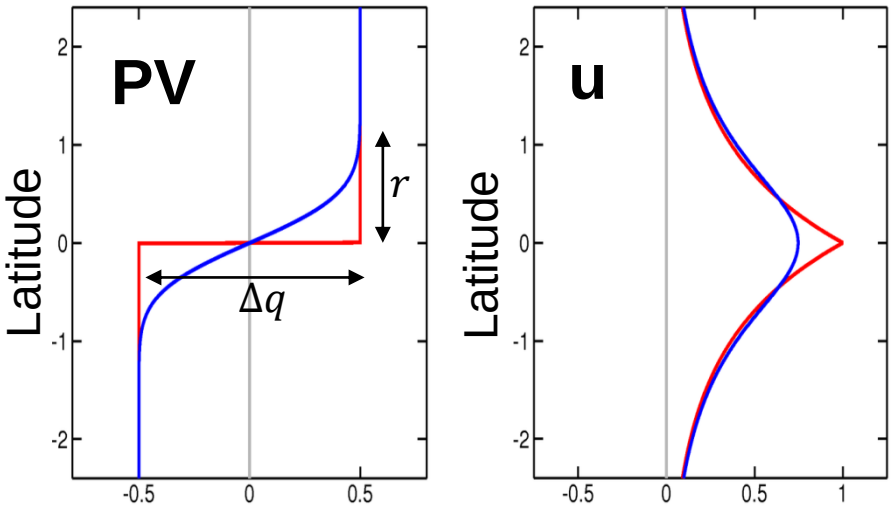
Jet maximum on PV gradient

Simplest single layer model

PV def: $q - f = \nabla^2\psi - \frac{\psi}{L_R^2} = \mathcal{L}(\psi)$

PV conservation: $\frac{Dq}{Dt} = 0$

PV inversion: $\psi = \mathcal{L}^{-1}(q - f)$

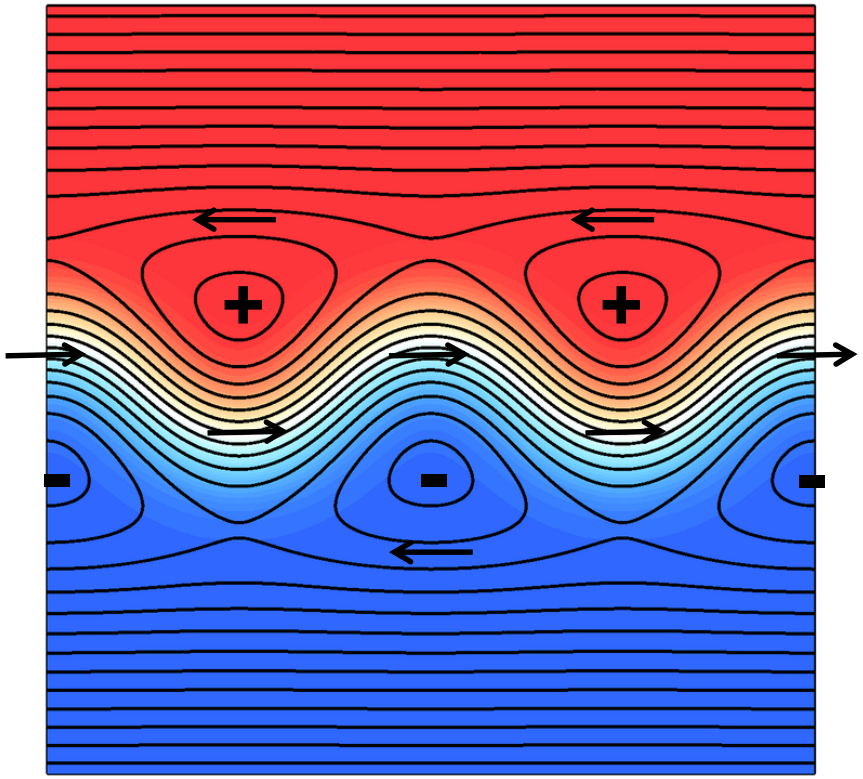


Meridional PV gradient (single layer QG)

$$\frac{\partial \bar{q}}{\partial y} = \beta - \frac{\partial}{\partial y}\left(\frac{\partial \bar{u}}{\partial y}\right)$$

Contours: $\psi_{stat} = \psi - cy$

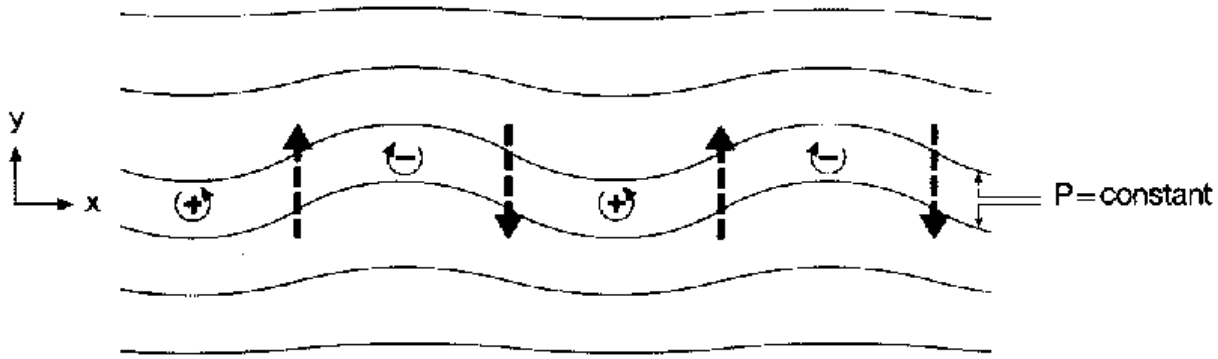
Shading: PV



meandering jet

Rossby Wave Propagation

Rossby waves propagate on horizontal PV gradients.



High vorticity

$\bar{q}_y > 0$

Low vorticity

Air displaced to south carries high vorticity and forms +ve q'

$\Rightarrow q'>0$ **induces** cyclonic circulation

$\Rightarrow$ advects air southwards on western flank

$\Rightarrow$ wave pattern propagates **westwards**

Phase speed: $c_p = \bar{u} - \frac{1}{k^2}\frac{\partial \bar{q}}{\partial y}$

Fig: Hoskins, McIntyre and Robertson (1985), *Q. J. Royal Met. Soc.*

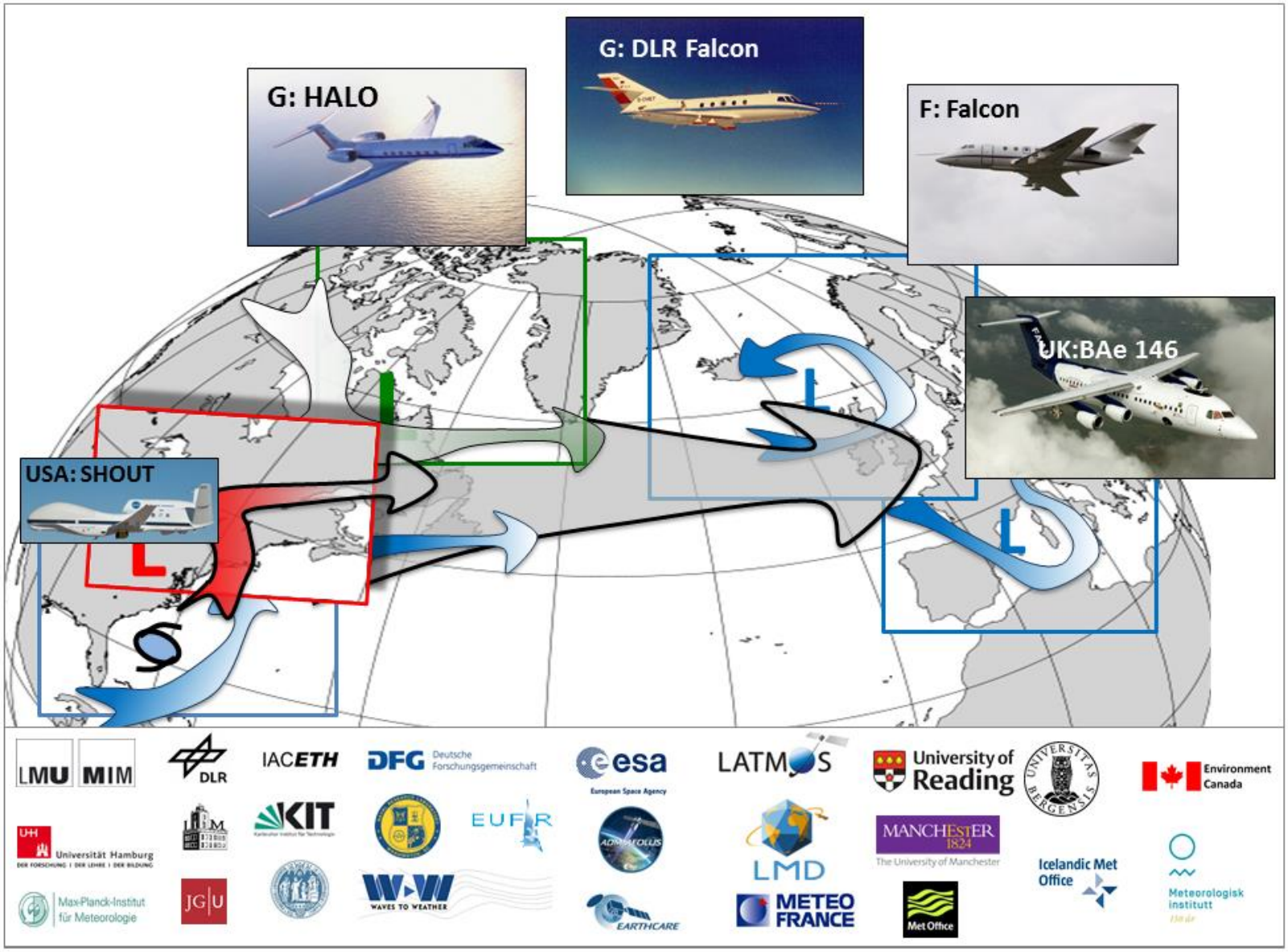
Remaining Talk Outline

1. **Does the jet stream structure differ from its representation in models (and analyses)?**

2. What are the ramifications for evolution of forecast errors?

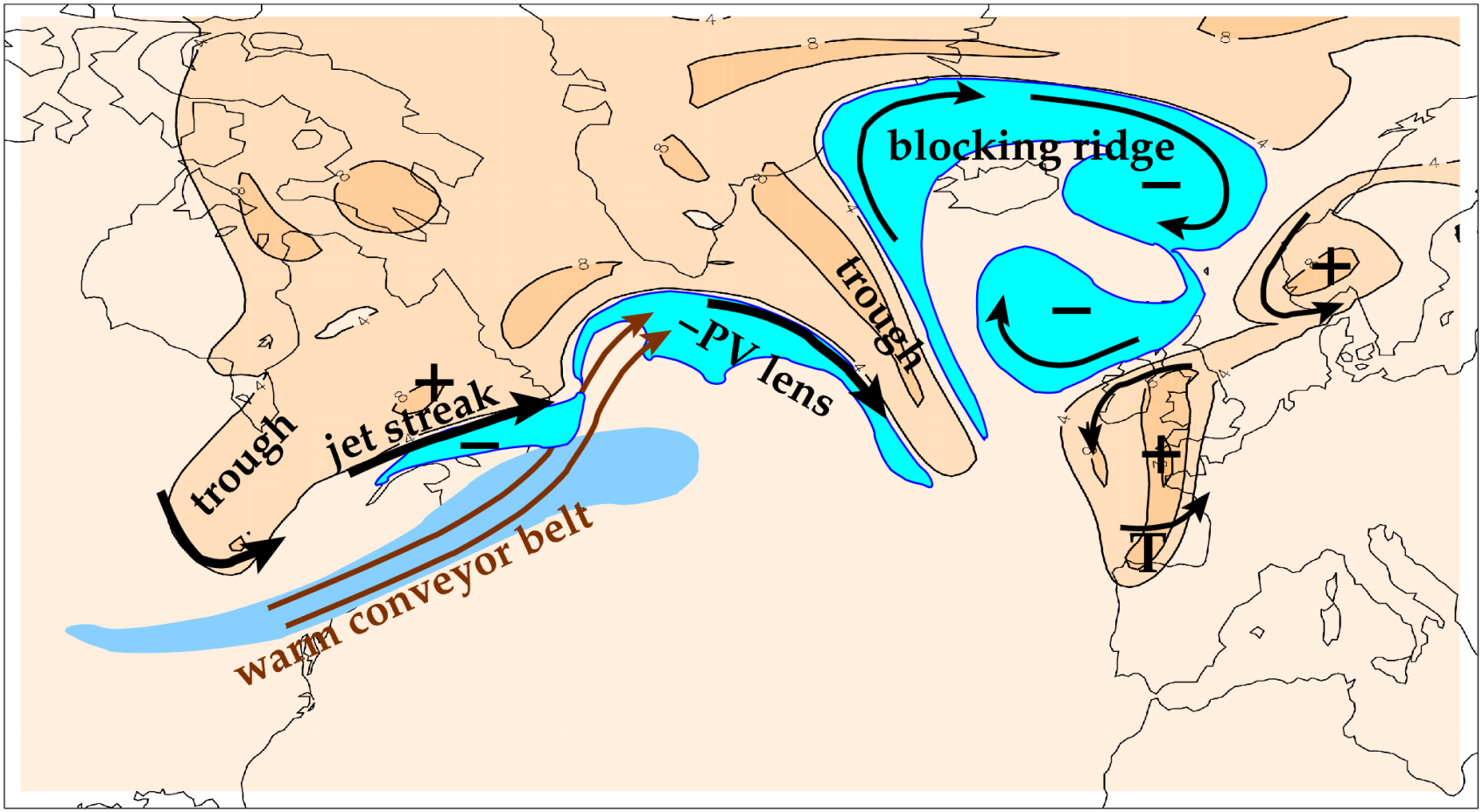
3. What are the chief mechanisms enabling diabatic influence on the dynamics of jet streams and Rossby waves?

North Atlantic Waveguide and Downstream Impacts Experiment (NAWDEX)



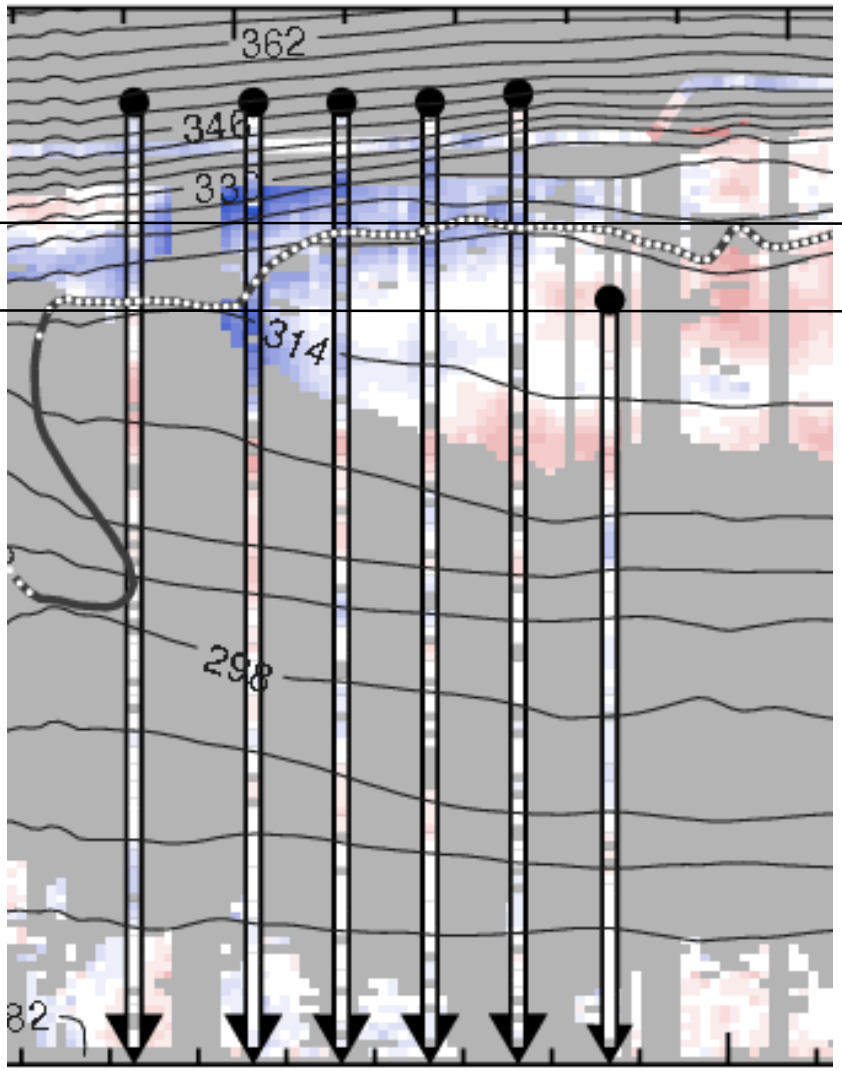
Overarching scientific aim of NAWDEX:

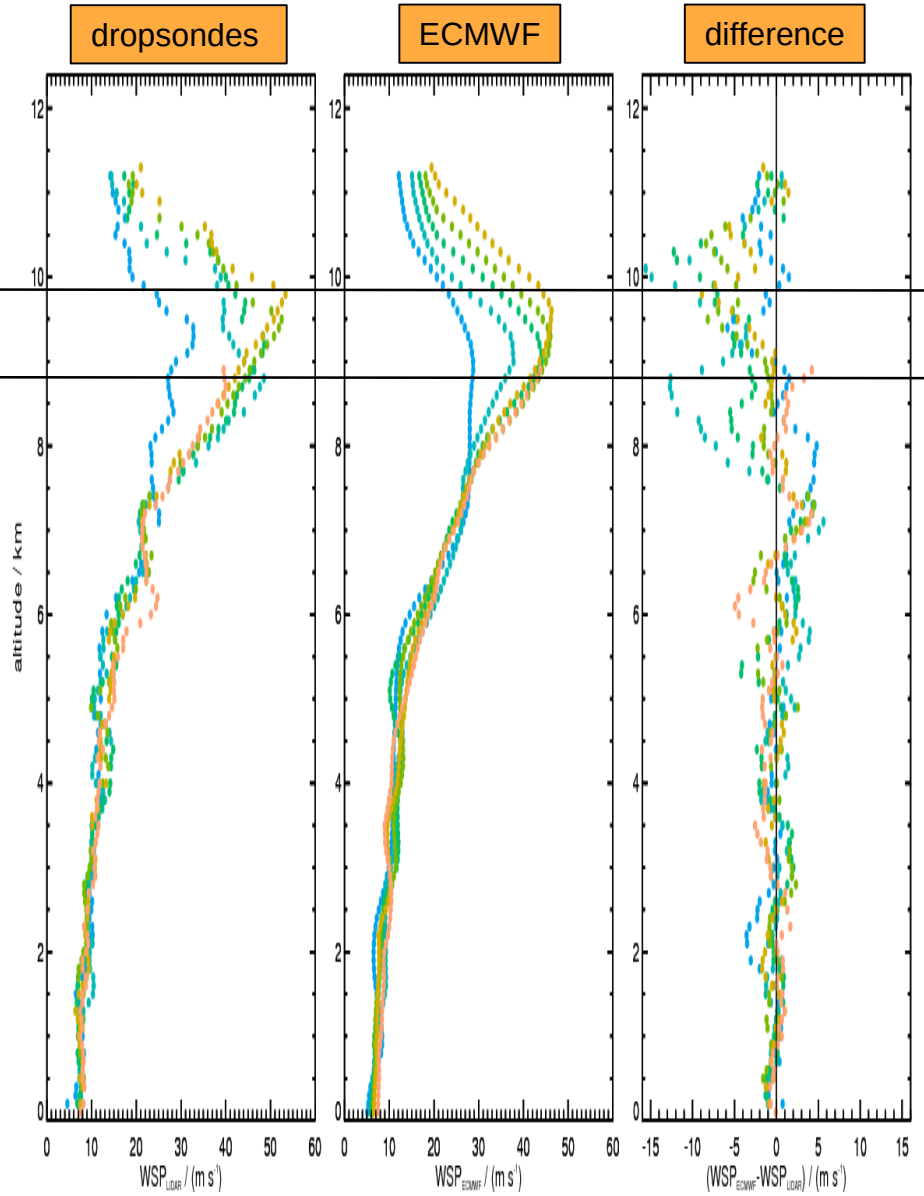
to quantify the effects of diabatic processes on disturbances to the jet stream near North America, their influence on downstream propagation across the North Atlantic, and consequences for high-impact weather in Europe.



Features related to the meandering tropopause and jet stream (orange is stratospheric air; cyan marks upper tropospheric PV anomalies).

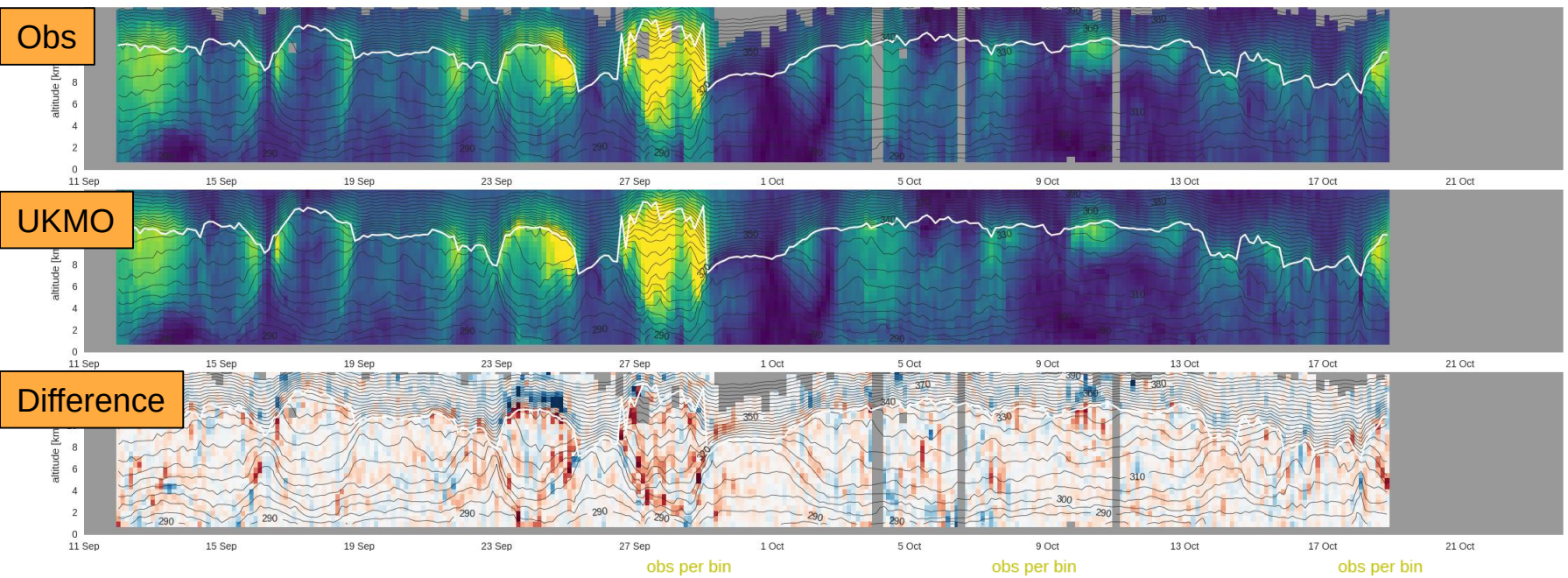
RF04: 23/9/16, wind speed: analysis – aircraft lidar obs

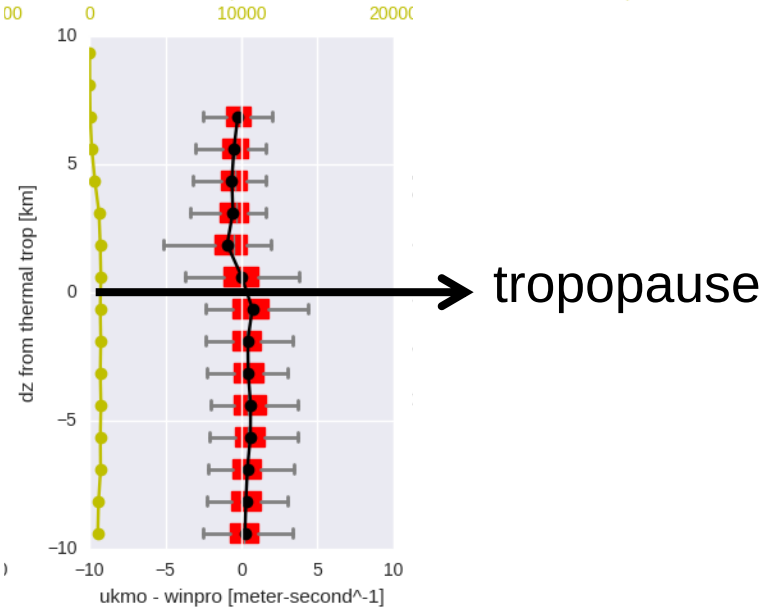




Obs vs model at T+0

Wind profiler @ South Uist

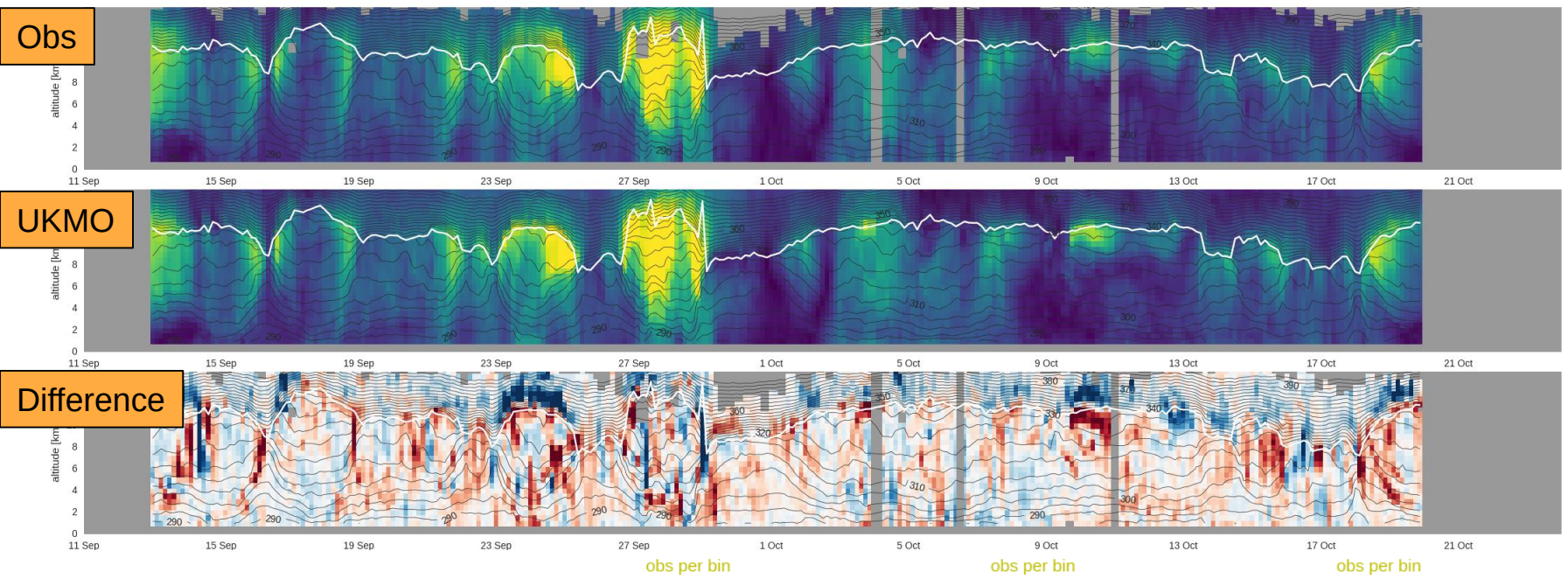


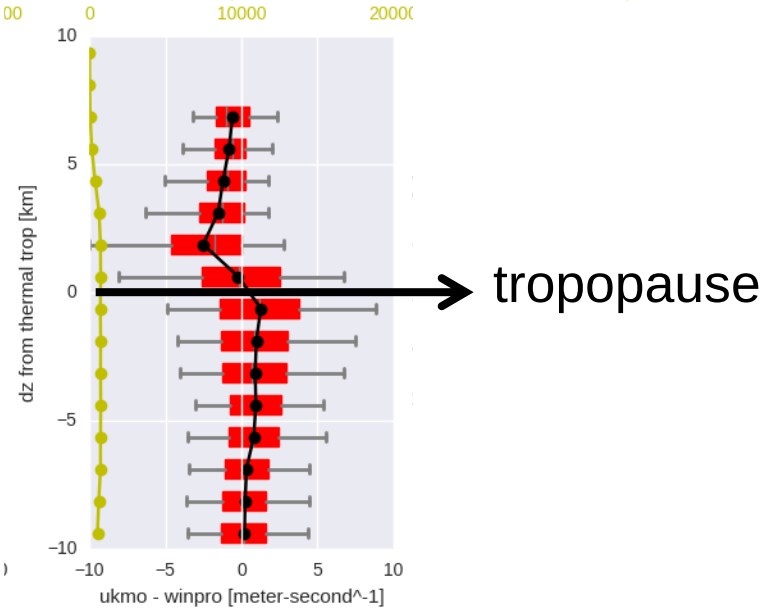


Wind profiler (ST radar) and analyses compared.

Composite relative to height of the tropopause on each profile.

Obs vs model at T+24





Wind profiler (ST radar) and analyses compared.

Composite relative to height of the tropopause on each profile.

Observed jet stream is "sharper"

2. What are ramifications for forecast evolution?

In vertical profiles, especially on the flanks of tropopause ridges, wind shear is observed to be much sharper than in analyses and forecasts.

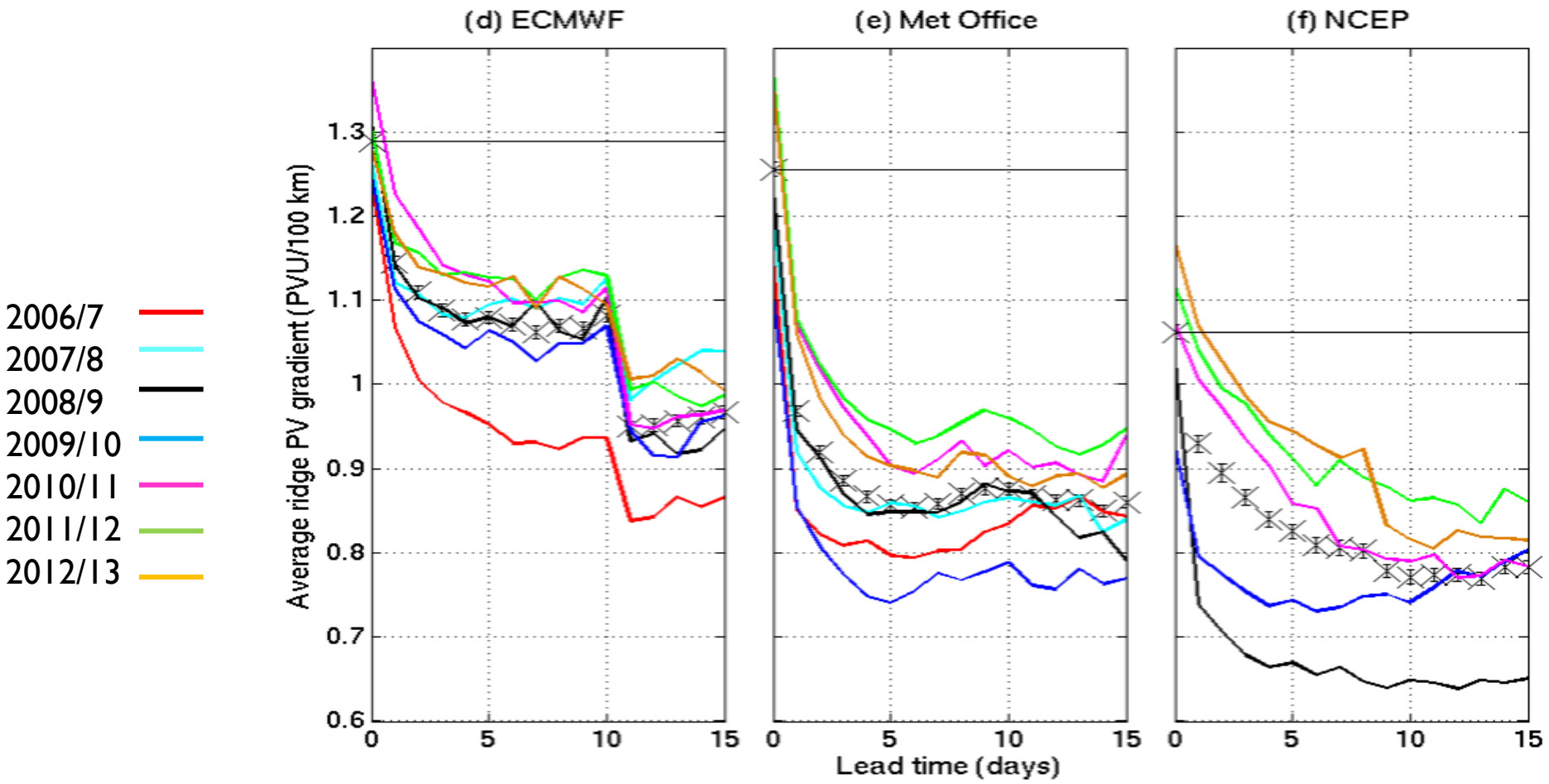
Under quasi-geostrophic approximation (for zonal flow):

Meridional PV gradient, $\frac{\partial \bar{q}}{\partial y} = \beta - \frac{\partial}{\partial y}\left(\frac{\partial \bar{u}}{\partial y}\right) - \frac{1}{\rho}\frac{\partial}{\partial z}\left(\rho \frac{f^2}{N^2}\frac{\partial \bar{u}}{\partial z}\right)$

Therefore, PV gradient must be too smooth in models

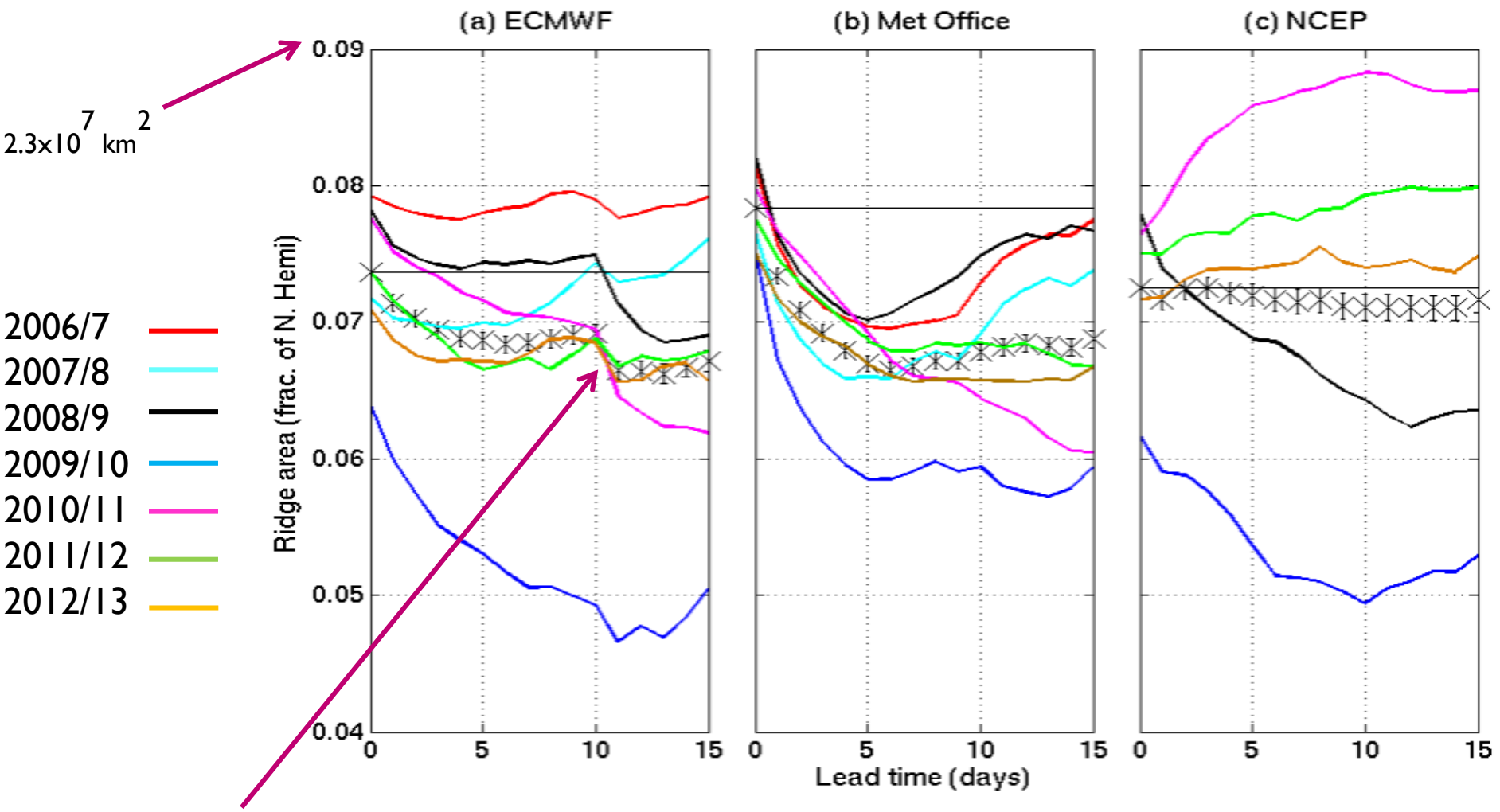
- How fast does PV gradient decrease with lead time?
- How does this affect Rossby waves?

Forecast of horizontal PV gradient across the tropopause (from TIGGE)



Gray, Dunning, Methven, Masato and Chagnon (2014), *GRL*

Forecasts of ridge area in Rossby waves



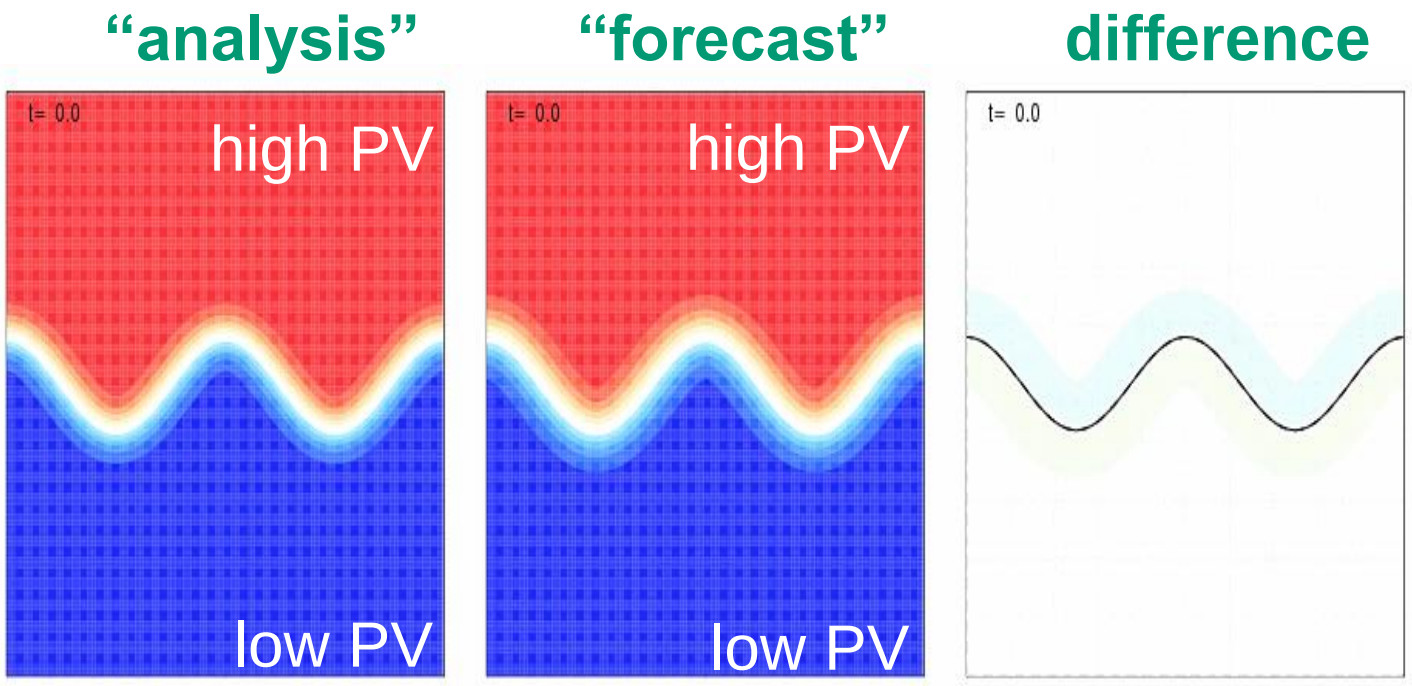
Reduced forecast resolution after day 10

Gray, Dunning, Methven, Masato and Chagnon (2014), *GRL*

Why does PV gradient sharpness matter?

- QG shallow water ("equivalent barotropic") equations:

$$\frac{Dq}{Dt} = 0 \qquad q = f_0 + \nabla^2\psi - \psi/L_R^2$$



$r_0 \approx 308\ km$ $r_0 \approx 381\ km$

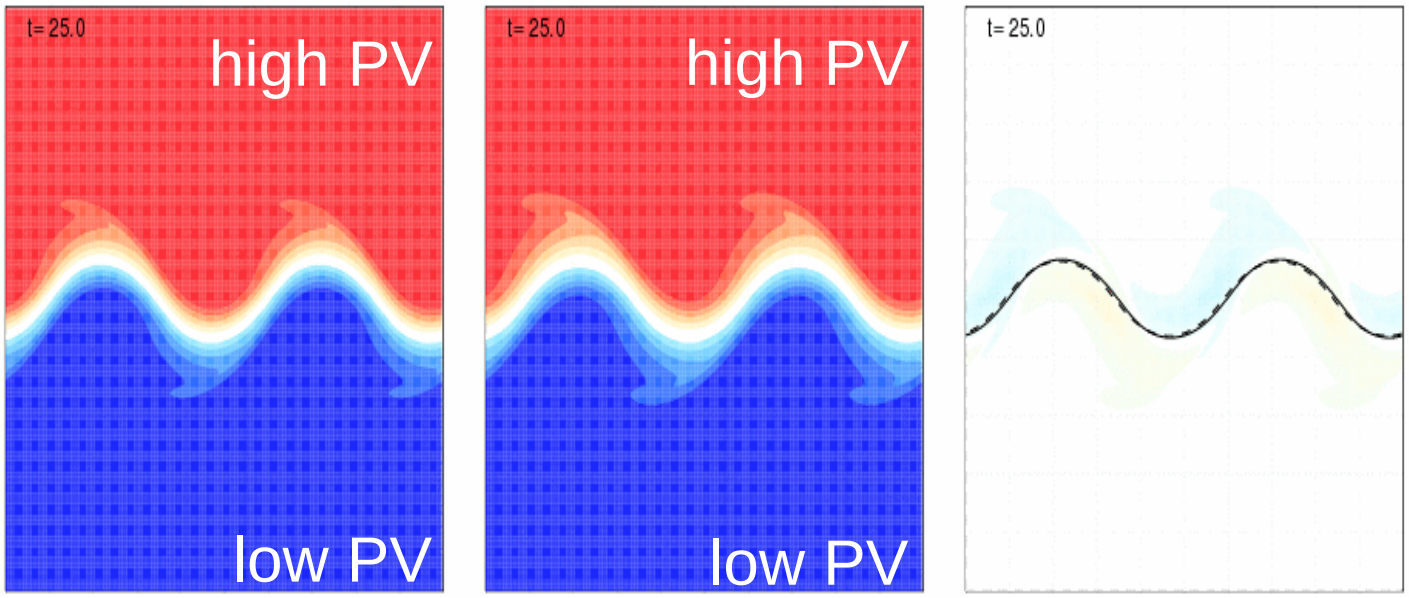
Use: $L_R = 700\ km$

A simple simulation

- QG shallow water ("equivalent barotropic") equations:

$$\frac{Dq}{Dt} = 0 \qquad q = f_0 + \nabla^2\psi - \psi/L_R^2$$

"analysis" "forecast" difference



$r_0 \approx 308\ km$ $r_0 \approx 381\ km$

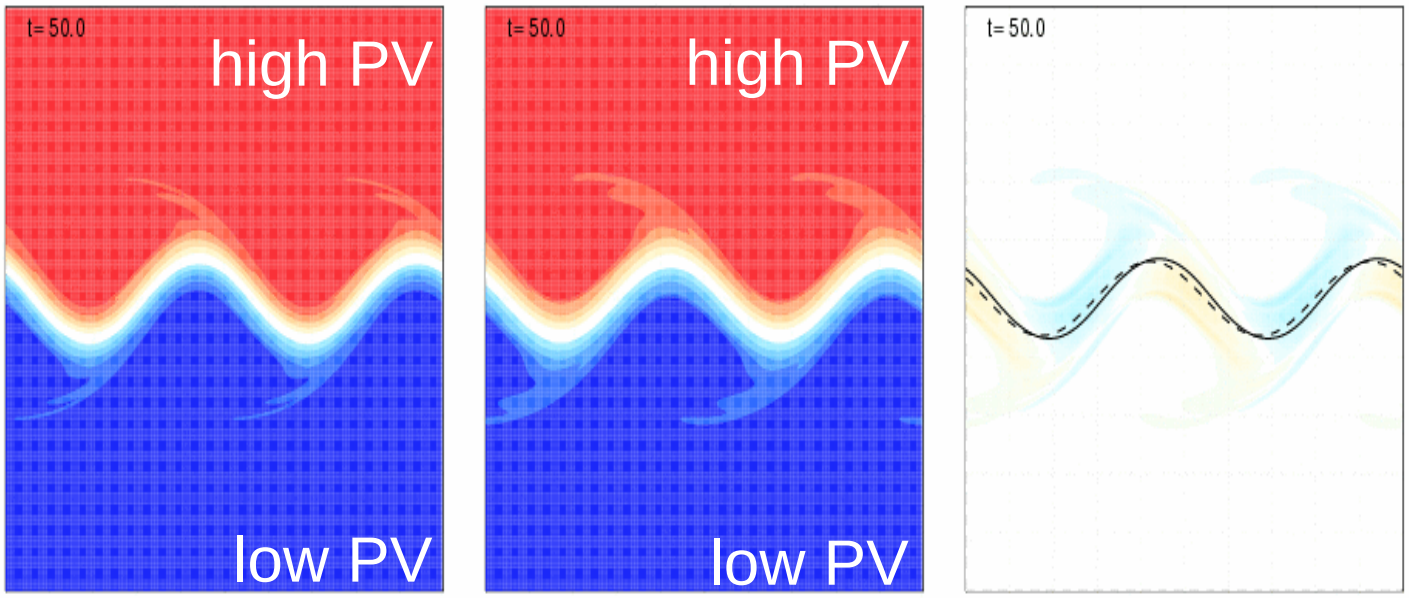
Use: $L_R = 700\ km$

A simple simulation

- QG shallow water ("equivalent barotropic") equations:

$$\frac{Dq}{Dt} = 0 \qquad q = f_0 + \nabla^2\psi - \psi/L_R^2$$

"analysis" "forecast" difference



$r_0 \approx 308\ km$ $r_0 \approx 381\ km$

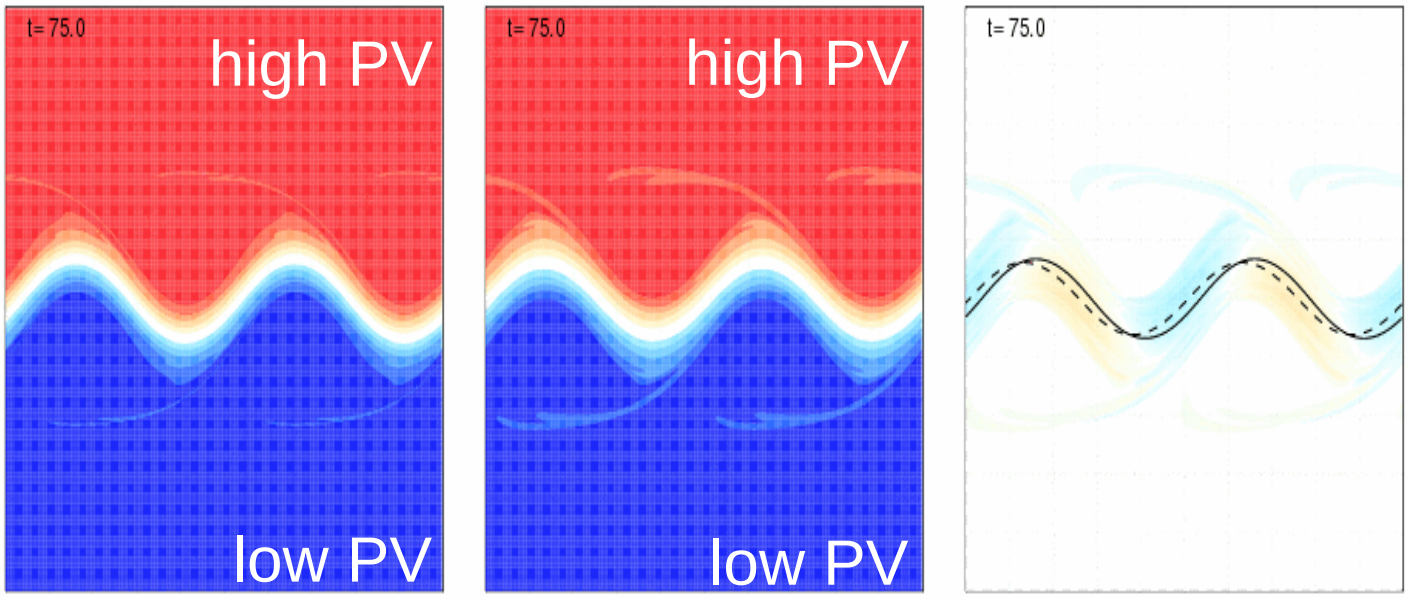
Use: $L_R = 700\ km$

A simple simulation

- QG shallow water ("equivalent barotropic") equations:

$$\frac{Dq}{Dt} = 0 \qquad q = f_0 + \nabla^2\psi - \psi/L_R^2$$

"analysis" "forecast" difference



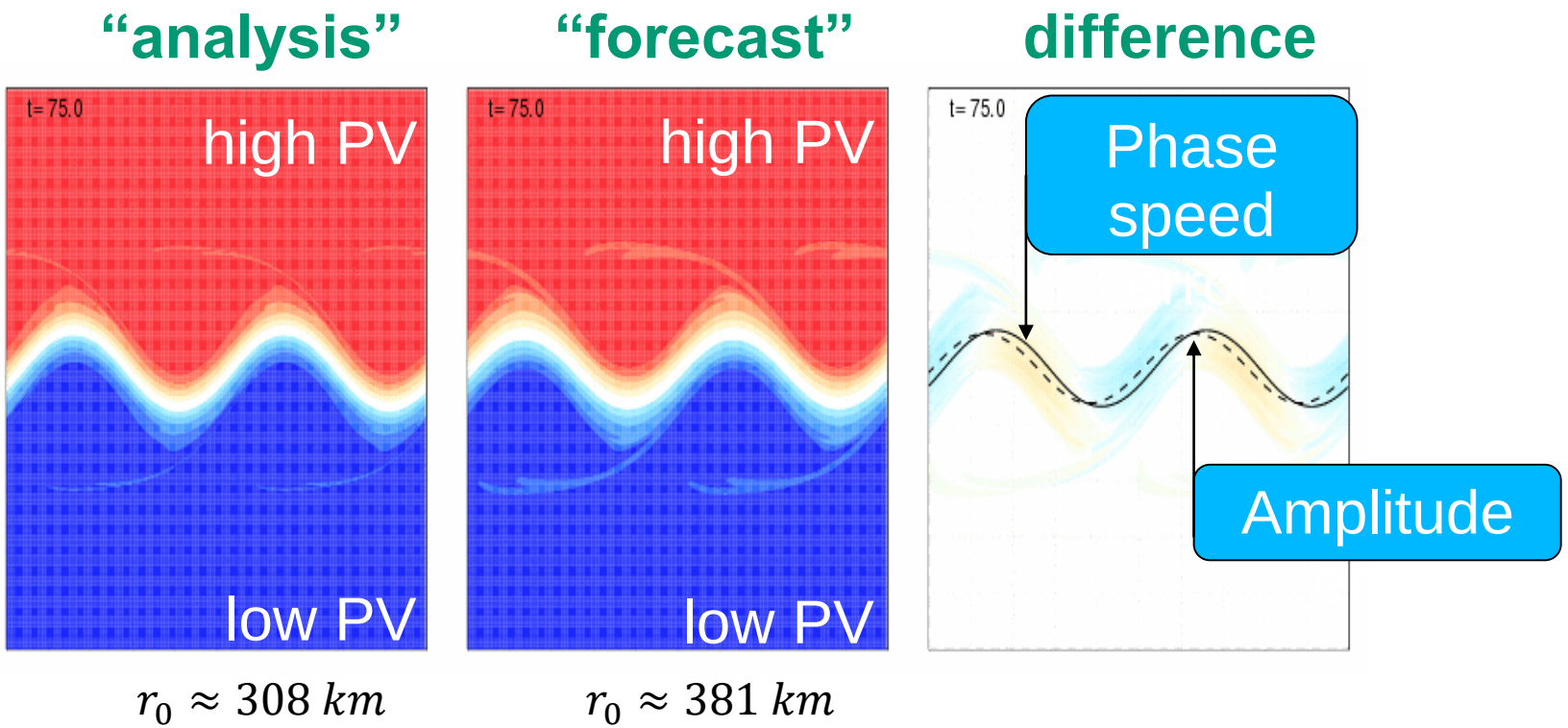
$r_0 \approx 308\ km$ $r_0 \approx 381\ km$

Use: $L_R = 700\ km$

A simple simulation

- QG shallow water ("equivalent barotropic") equations:

$$\frac{Dq}{Dt} = 0 \qquad q = f_0 + \nabla^2\psi - \psi/L_R^2$$



Use: $L_R = 700\ km$

RW propagation on smooth PV step

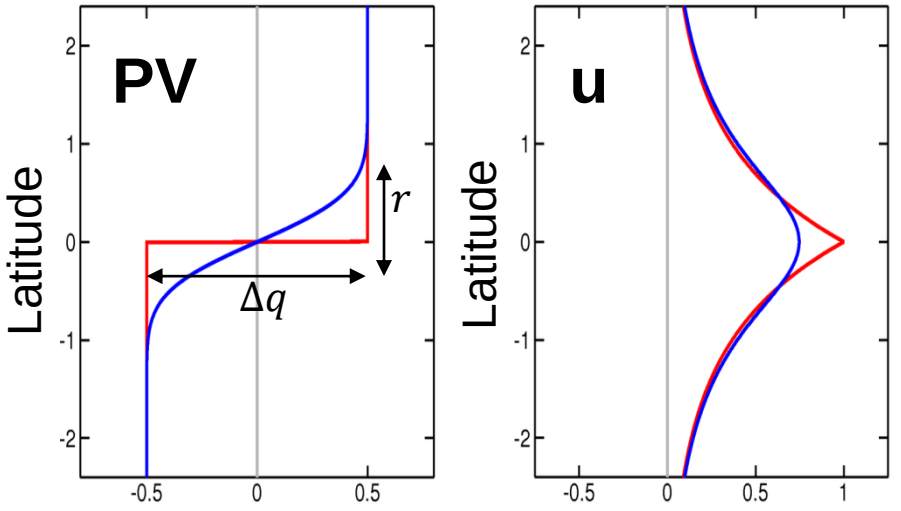
Single step in PV is smoothed by convolution with a weight.

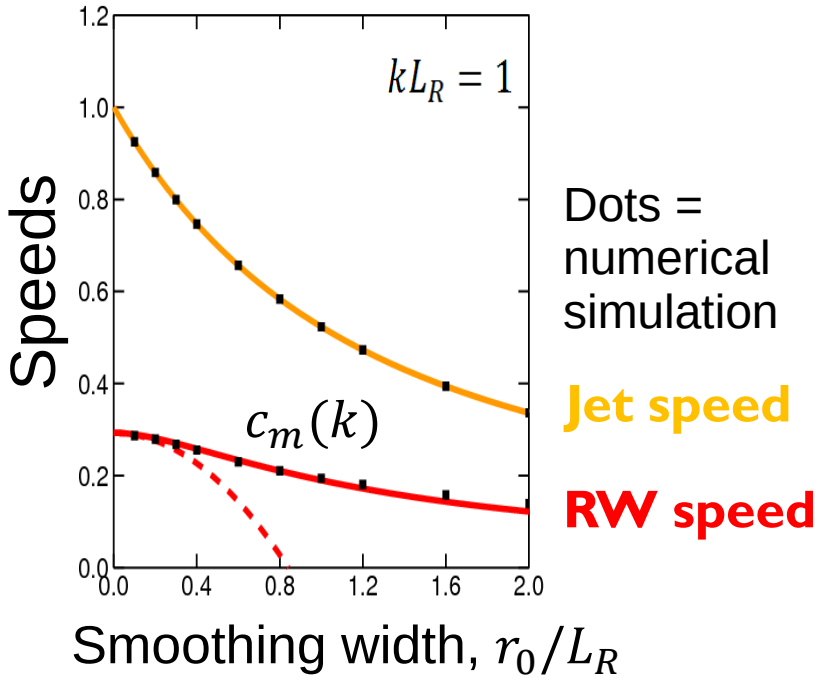
Solution obtained in limit where tropopause width is narrow compared with wavelength ($kr_0 \ll 1$)

General conclusion: smoothing PV gradient reduces both jet maximum and counter-propagation rate …
but jet decrease is greater

⇒ phase speed reduces

⇒ affects shorter waves more, increasing dispersion





Dots = numerical simulation

Jet speed

RW speed

Harvey *et al*, *J. F. Mech.* (2016)

Decrease of jet meander amplitude

- Fast oscillation:

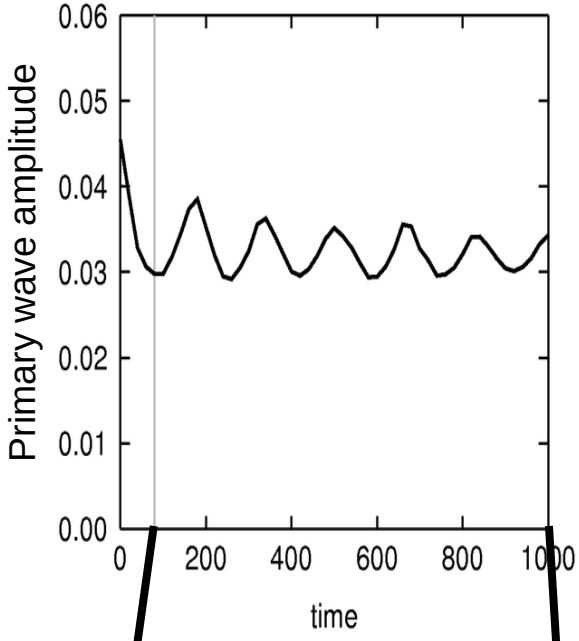
  through advection of PV filaments around the cat's eyes on jet flanks

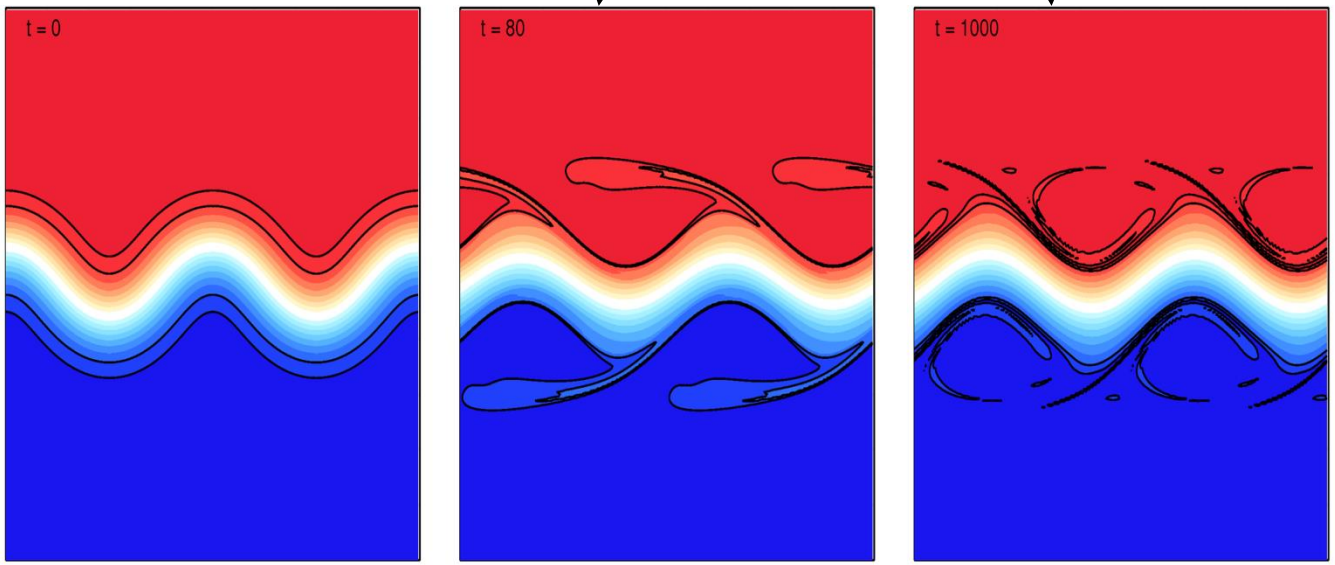
- Gradual decrease:

  mixing of PV within the critical layer

- Wave activity fluxes into jet flanks, but **global WA is conserved**

  $\Rightarrow$ WA of jet meander must decrease.





Harvey *et al*,
J. Atmos. Sci. (2018)

3. Mechanisms for diabatic influence

Focus on the diabatic influences on Rossby waves at tropopause level (*wave guide disturbances*).

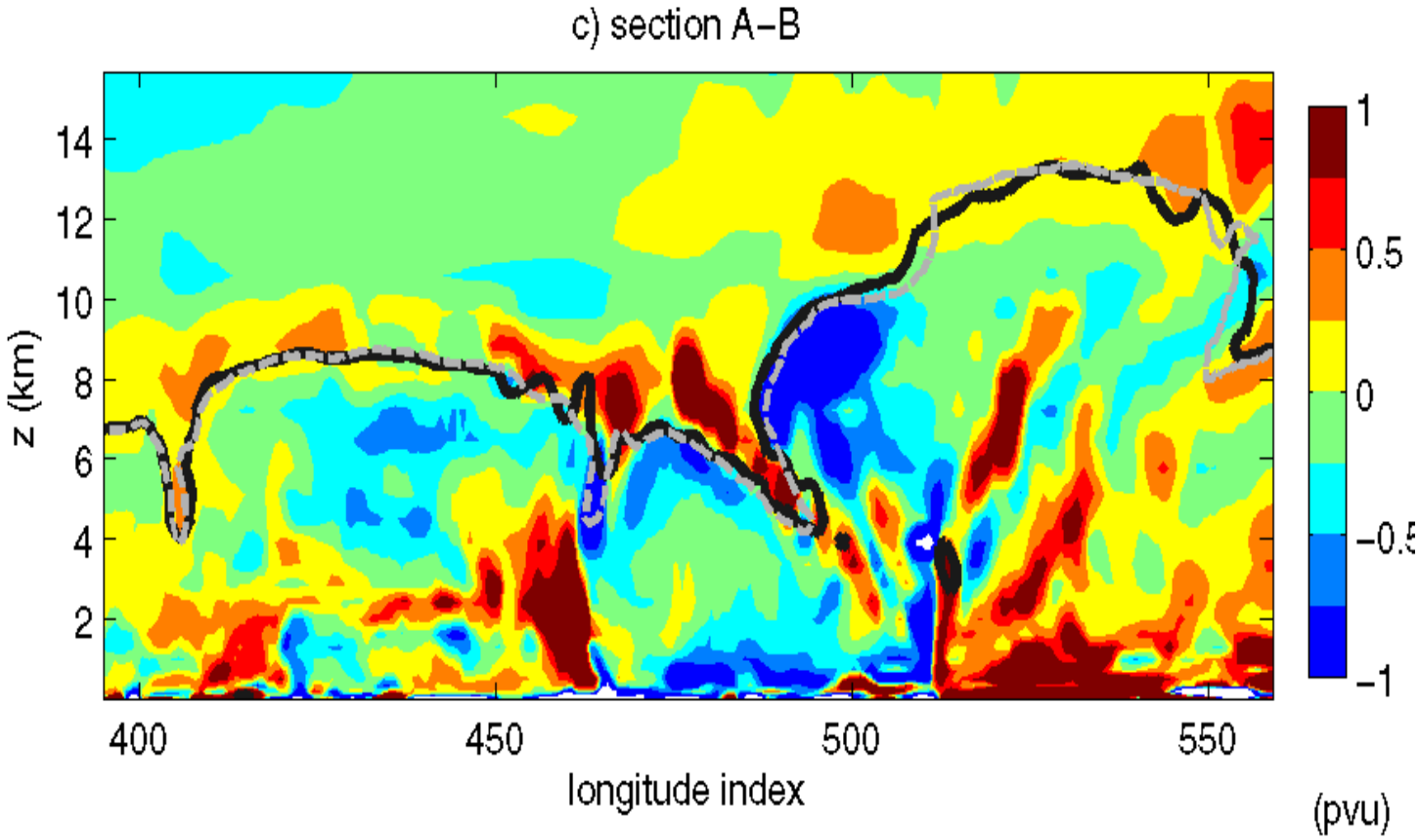
A. Diabatic maintenance of PV contrast across the tropopause

B. Sharpening of jet stream PV gradient and max winds by "non-advective PV flux" (heating in vertical wind shear)

C. Amplification of baroclinic wave growth rate, through lower "effective static stability" ***(Heini Wernli)***

D. Diabatic mass transport into upper tropospheric ridges and importance for mid-latitude blocking ***(Christian Grams)***

A. Radiative maintenance of PV contrast across tropopause

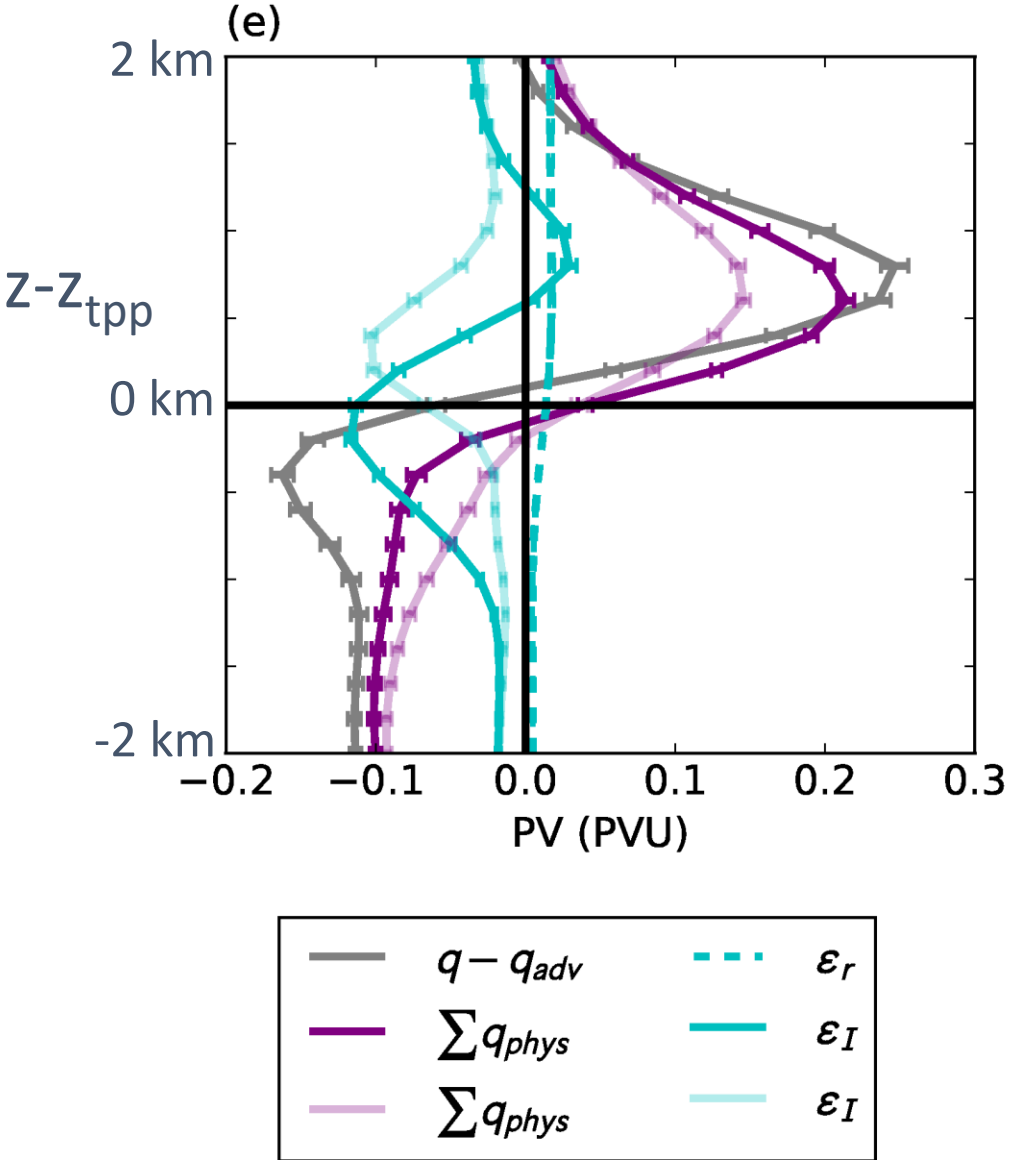


- Positive *diabatic PV* above (on strat side) of tropopause
- Negative *diabatic PV* beneath (on trop side) of tropopause
- Tropopause elevation not significantly altered by **direct diabatic PV modification**

Chagnon, Gray and Methven (2013), *Q J R Met S*

Composite of 92 forecasts

Relative to the tropopause in troughs



Troughs defined by locations where

$$\theta'(\lambda, \phi, PV2, t) < 0$$

Average on set of levels defined by vertical distance from the tropopause

Physics parameterisations reduce PV below tropopause (and increase it above)

Dynamical core of Met Office Unified Model reduces PV at tropopause

Saffin *et al* (2017), *JGR*

Ramifications for synoptic scale evolution

PV conservation equation …….. and PV-tracer

$$\frac{Dq}{Dt} = \frac{\partial q}{\partial t} + \underline{u}.\nabla q = \sum_i S_i \qquad q_p(0) = q(0) \qquad \frac{Dq_p}{Dt} = 0$$

Using observation of little *direct diabatic PV modification* at tropopause:

$$q_i(tpp) \approx 0 \qquad q(tpp) \approx q_p$$

However, diabatic effects are felt mainly through induced wind, $\underline{u}_i$

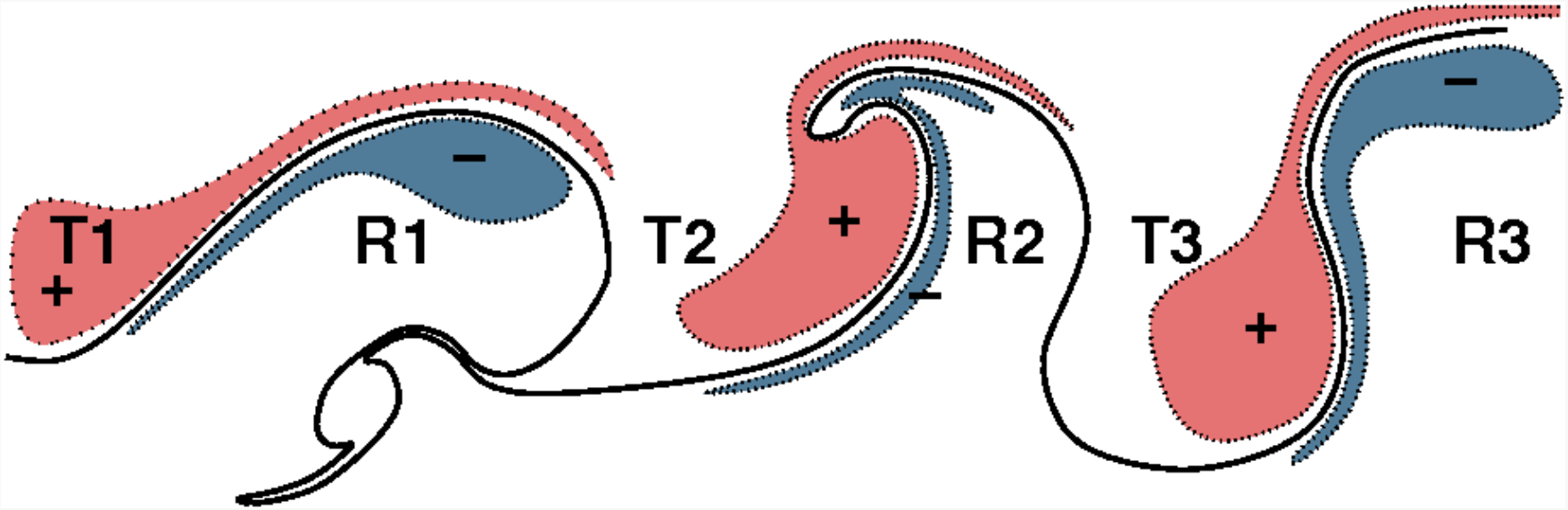
$$\frac{\partial q_p}{\partial t} + (\underline{u}_p + \sum_i \underline{u}_i).\nabla q_p \approx 0 \qquad \underline{u}_i = L^{-1}(q_i)$$

Indirect PV modification by diabatic processes via changes to winds.

Links directly with moist baroclinic wave theory (De Vries, *2010, JAS*).

Cooperative coupling between diabatic PV in troposphere and Rossby waves along tropopause or lower boundary.

Diabatic influence on Rossby waves



1. LW cooling max at tropopause (humidity step)
   - ⇒ "diabatic PV dipole" with ***little direct PV change*** at tropopause
   - ⇒ but, enhances tropopause PV gradient
   - ⇒ would be influenced by cirrus just under tropopause
2. Diabatic PV enhances PV anomaly pattern of Rossby wave
   - ⇒ greater westward propagation and enhanced baroclinic interaction
   - ⇒ But, also stronger jet associated with greater PV contrast

Chagnon, Gray and Methven (2013), *QJRMetS*

B. Jet sharpening by non-advective PV flux

Alternative Lagrangian form of PV equation.

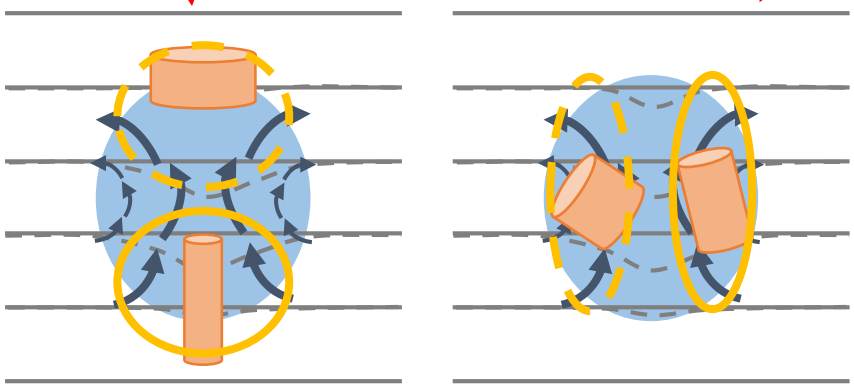
Evolution following flow within an isentropic layer

$$\frac{\tilde{D}}{Dt} = \frac{\partial}{\partial t} + \tilde{\boldsymbol{u}} \cdot \nabla$$

$$\rho \frac{\tilde{D}P}{Dt} = P\nabla \cdot (\rho \boldsymbol{u}_J) + \nabla \cdot \left(\boldsymbol{\zeta}_{//}\dot{\theta} + \boldsymbol{F} \times \nabla\theta\right)$$

Concentration/dilution of PV below/above heating

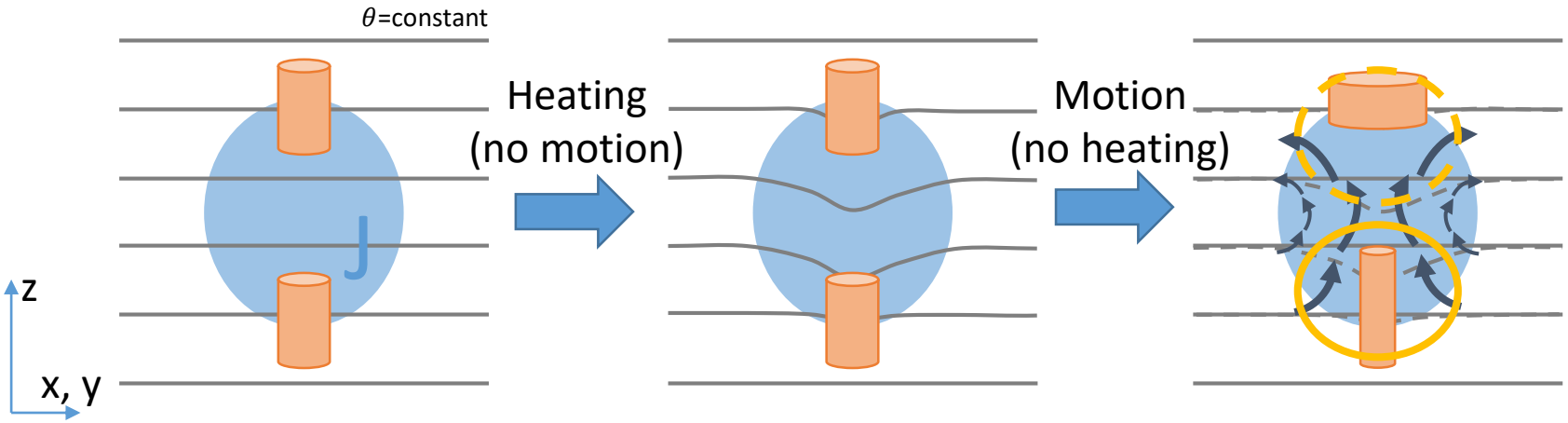
Non-advective transport of PV along θ surfaces



Negative PV cannot arise through diabatic mass flux convergence

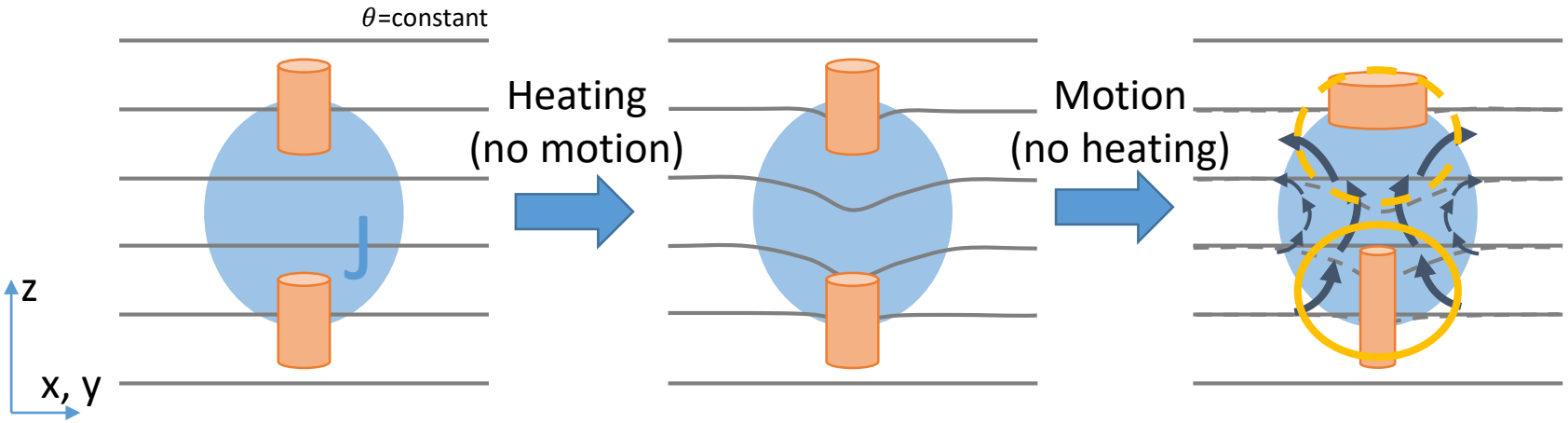
If P>0 initially, it must remain positive through the first term on right.

Thought experiment



- $\rho P = \boldsymbol{\zeta} \cdot \nabla\theta$

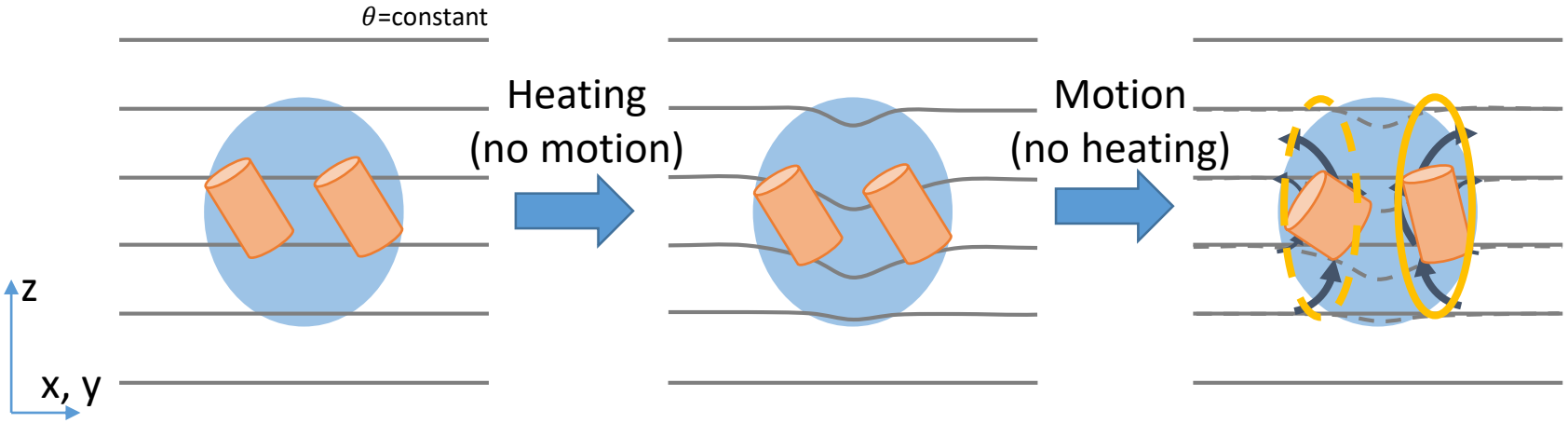
Thought experiment



- $\rho P = \boldsymbol{\zeta} \cdot \nabla\theta$

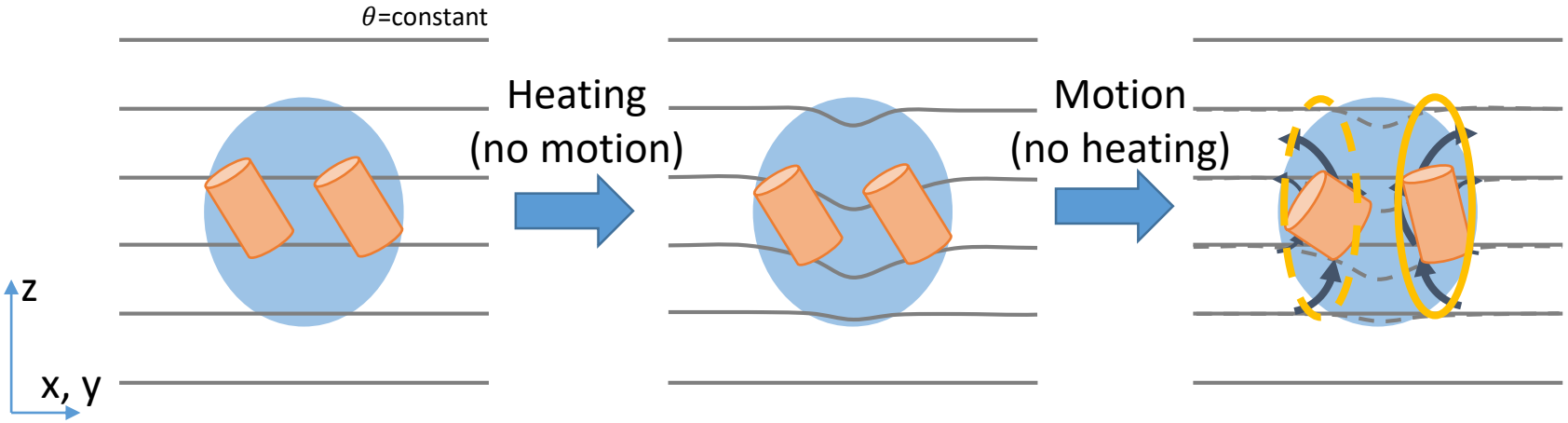
- Vortex stretching induced by adjustment to heating
  → vertical PV dipole
- But…negative PV cannot be produced
  (only a PV reduction towards zero)

Thought experiment



Now include some horizontal vorticity

Thought experiment



Now include some horizontal vorticity

- Vortex tilting induced by response to heating
  → horizontal PV dipole
- Requires a horizontal component of absolute vorticity (= vertical wind shear)
- No constraint on sign of PV values (negative PV can be produced)

Response to “WCB” heating (imposed)

2-D semi-geostrophic model with heating:

$$\frac{Du_g}{Dt} - f v_{ag} = 0$$

$$\frac{D\theta}{Dt} = H$$

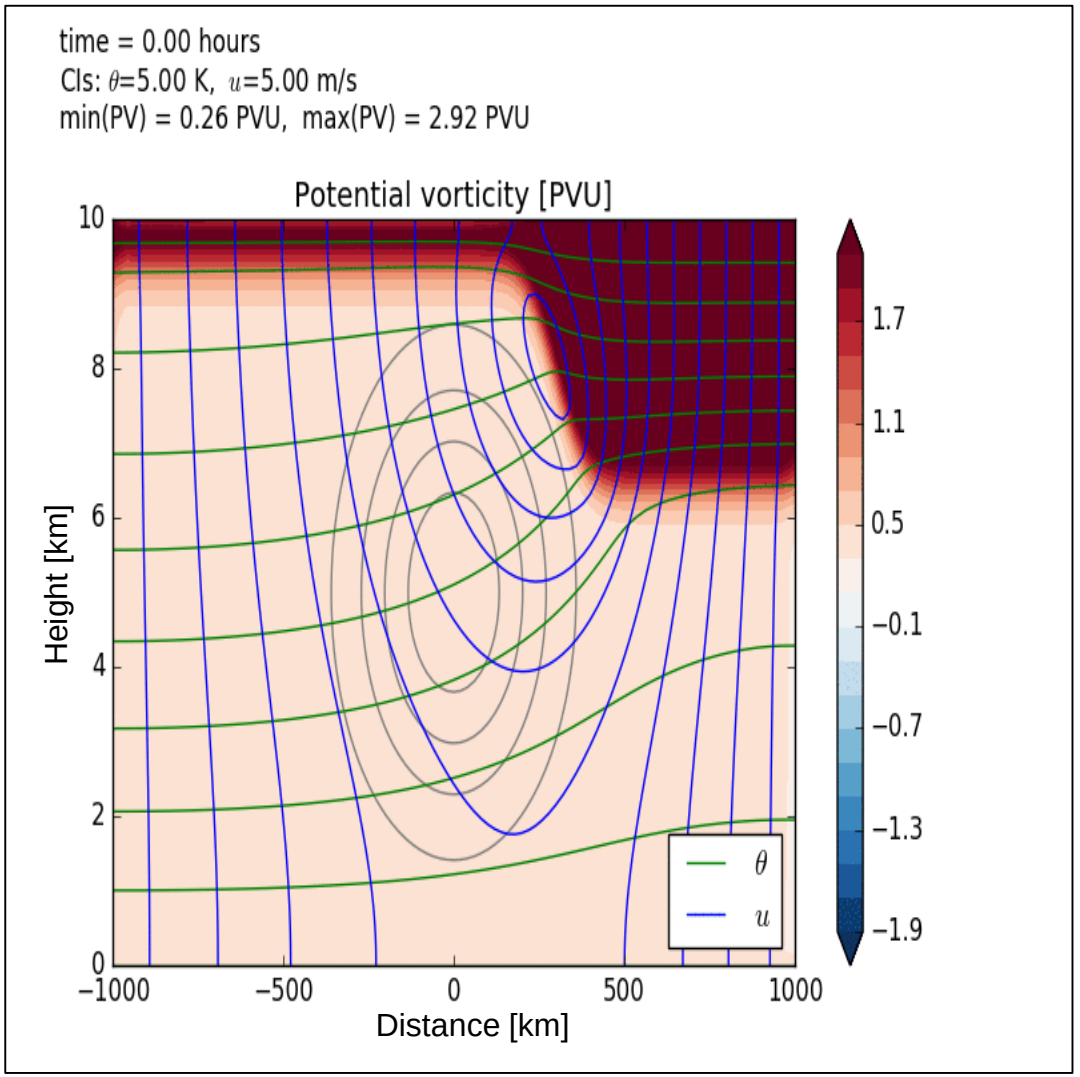
$$\frac{D}{Dt} = \frac{\partial}{\partial t} + \left(u_g + u_{ag}\right)\frac{\partial}{\partial x} + v_{ag}\frac{\partial}{\partial y} + w\frac{\partial}{\partial z}$$

Define cross-frontal streamfunction, ψ

$$v_{ag} = -\frac{\partial \psi}{\partial z} \qquad w = \frac{\partial \psi}{\partial y}$$

Meridional circulation acts to maintain TWB:

$$\psi_{yy}\theta_z + \psi_{yz}\left(u_z - \theta_y\right) + \psi_{zz}\left(f - u_y\right) = H_y$$



Response to “WCB” heating (imposed)

2-D semi-geostrophic model with heating:

$$\frac{Du_g}{Dt} - f v_{ag} = 0$$

$$\frac{D\theta}{Dt} = H$$

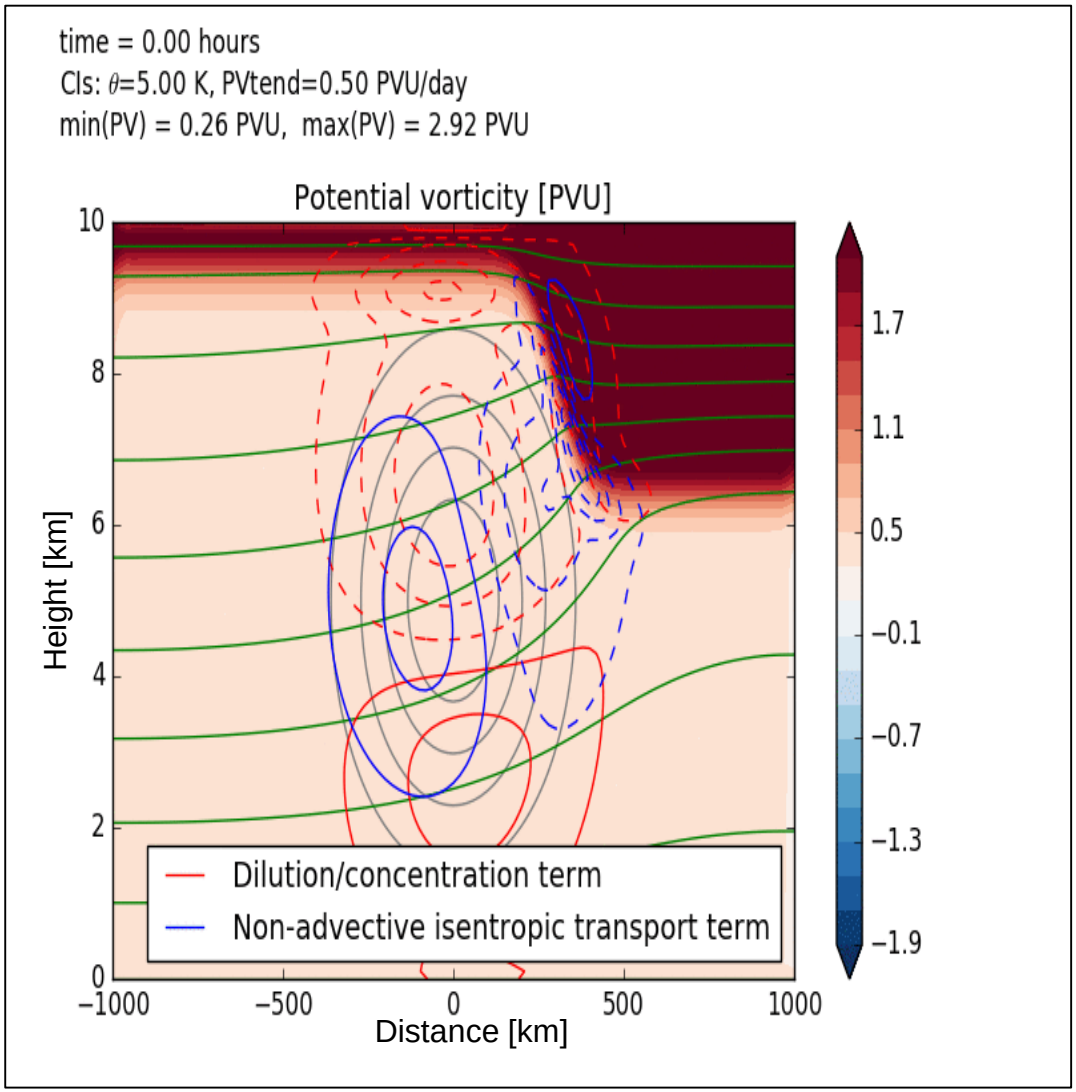
$$\frac{D}{Dt} = \frac{\partial}{\partial t} + \left(u_g + u_{ag}\right)\frac{\partial}{\partial x} + v_{ag}\frac{\partial}{\partial y} + w\frac{\partial}{\partial z}$$

Define cross-frontal streamfunction, ψ

$$v_{ag} = -\frac{\partial \psi}{\partial z} \qquad w = \frac{\partial \psi}{\partial y}$$

Meridional circulation acts to maintain TWB:

$$\psi_{yy}\theta_z + \psi_{yz}\left(u_z - \theta_y\right) + \psi_{zz}\left(f - u_y\right) = H_y$$



Response to “WCB” heating (imposed)

2-D semi-geostrophic model with heating:

$$\frac{Du_g}{Dt} - f v_{ag} = 0$$

$$\frac{D\theta}{Dt} = H$$

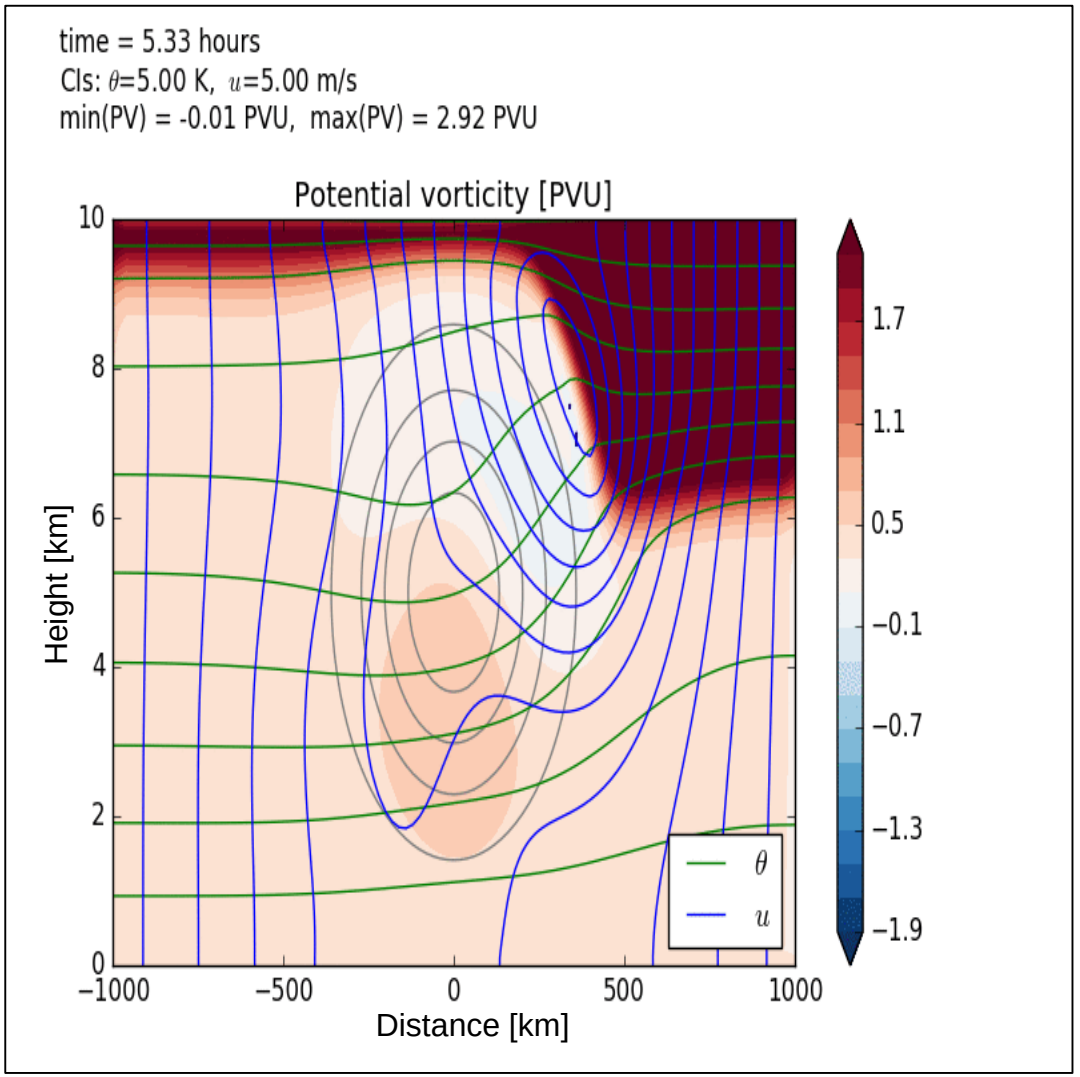
$$\frac{D}{Dt} = \frac{\partial}{\partial t} + \left(u_g + u_{ag}\right)\frac{\partial}{\partial x} + v_{ag}\frac{\partial}{\partial y} + w\frac{\partial}{\partial z}$$

Define cross-frontal streamfunction, ψ

$$v_{ag} = -\frac{\partial \psi}{\partial z} \qquad w = \frac{\partial \psi}{\partial y}$$

Meridional circulation acts to maintain TWB:

$$\psi_{yy}\theta_z + \psi_{yz}\left(u_z - \theta_y\right) + \psi_{zz}\left(f - u_y\right) = H_y$$



Conclusions

1. **PV gradient at tropopause is too weak** in global forecasts and declines

   ⇒ expect Rossby waves to move eastwards too slowly in forecasts

   ⇒ Rossby wave amplitude declines (on average)

   Mixing too strong, or diabatic maintenance too weak?

2. **Negative PV** observed is a signature of diabatic "sharpening" of jet stream
   - Via *non-advective PV flux* which requires heating in vertical wind shear

3. **Effects of heating in large-scale vertical wind shear are systematic,** but model error may arise from heating profile and position in shear flow *(recall Mark Rodwell's talk)*

Conclusions

1. **PV gradient at tropopause is too weak** in global forecasts and declines

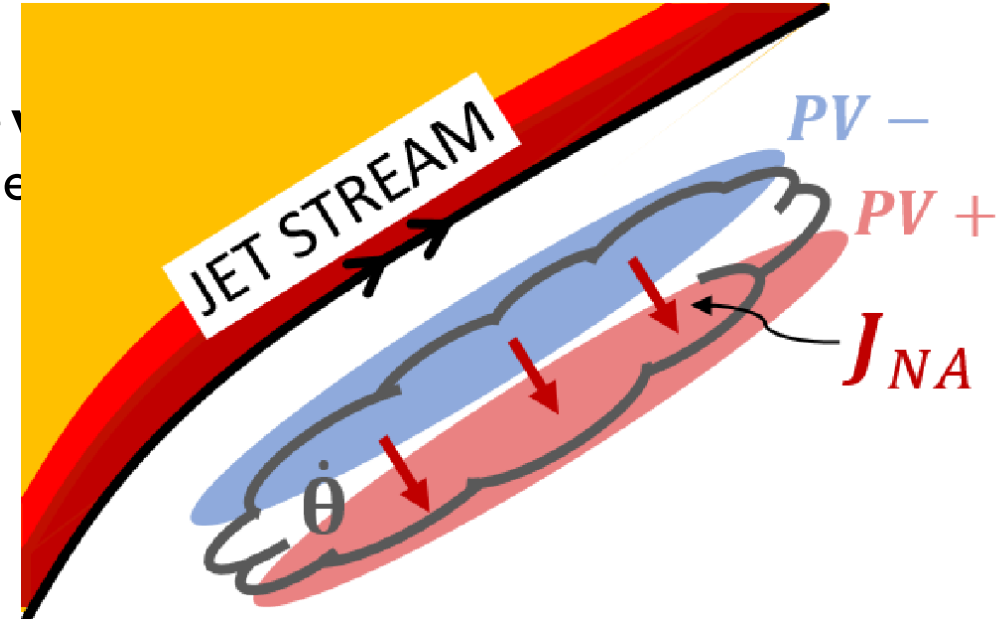
   ⇒ expect Rossby waves to move eastwards too slowly in forecasts

   ⇒ Rossby wave amplitude declines (on average)

   Mixing too strong, or diabatic maintenance too weak?

2. **Negative PV** observed is a signature of diabatic "sharpening" of jet stream
   - Via *non-advective PV flux* which requires heating in vertical wind shear

3. **Effects of heating in large-scale v**
   but model error may arise from he
   flow *(recall Mark Rodwell's talk)*



Thankyou for your attention