Mechanisms for Diabatic Influence on the Jet Stream

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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

\[
P = \frac{1}{\rho} \zeta \cdot \nabla \theta = (\zeta \cdot n) \frac{|\nabla \theta|}{\rho}
\]

where

\[
n = \frac{\nabla \theta}{|\nabla \theta|}
\]
Tracking non-conservative changes in PV

- PV tracer diagnostics are based on a Lagrangian decomposition of the PV field.
- Full PV conservation equation can be written:
  \[
  \frac{Dq}{Dt} = \sum_i S_i \quad \text{where } S_i \text{ denotes the Lagrangian tendency resulting from one physical process in model}
  \]

- **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.
  \[q = q_{\text{passive}} + \sum_i q_i + \text{error}\]
  \[
  \begin{align*}
  q_i(t_0) &= 0 \\
  q_{\text{passive}}(t_0) &= q(t_0)
  \end{align*}
  \]
PV along a section through jet stream and WCB

(a) Full PV

(b) Advected initial PV

(c) PV change due to all physics

(d) PV change due to dyn. core
PV along a section through jet stream and WCB

(a) Full PV

(b) PV change due to all physics

(c) Cross-jet wind shear along HALO flight track

- In situ: raw
- In situ: 600s mean
- MetUM
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: \( \psi = \mathcal{L}^{-1}(q - f) \)

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**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

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meandering jet
Rossby waves propagate on horizontal PV gradients.

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:

$$c_p = \frac{-u}{k^2} - \frac{1}{\partial_y q}$$

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
Wind profiler (ST radar) and analyses compared.

Composite relative to height of the tropopause on each profile.
Wind profiler (ST radar) and analyses compared.

Composite relative to height of the tropopause on each profile.
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, \[ \frac{\partial \tilde{q}}{\partial y} = \beta - \frac{\partial}{\partial y} \left( \frac{\partial \tilde{u}}{\partial y} \right) - \frac{1}{\rho} \frac{\partial}{\partial z} \left( \rho \frac{f^2}{N^2} \frac{\partial \tilde{u}}{\partial z} \right) \]

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

2.3x10^7 km^2


(a) ECMWF

(b) Met Office

(c) NCEP

Ridge area (frac. of N. Hem)

Lead time (days)

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:

\[
\frac{Dq}{Dt} = 0 \quad q = f_0 + \nabla^2 \psi - \psi / L_R^2
\]

Use: \( L_R = 700 \, \text{km} \)
A simple simulation

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\frac{Dq}{Dt} = 0 \\
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\]

"analysis" "forecast" difference

- High PV

- Low PV

- \( r_0 \approx 308 \text{ km} \)

- \( r_0 \approx 381 \text{ km} \)

Use: \( L_R = 700 \text{ km} \)
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A simple simulation

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

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\frac{Dq}{Dt} = 0 \\
q = f_0 + \nabla^2 \psi - \psi / L_R^2
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"analysis"  "forecast"  difference

high PV  low PV  high PV  low PV

\[r_0 \approx 308 \text{ km} \quad r_0 \approx 381 \text{ km} \]

Use: \(L_R = 700 \text{ km}\)
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

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
  ⇒ WA of jet meander must decrease.

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

\[
\frac{Dq}{Dt} = \frac{\partial q}{\partial t} + u_i \nabla q = \sum_i S_i \quad q_p(0) = q(0) \quad \frac{Dq_p}{Dt} = 0
\]

Using observation of little direct diabatic PV modification at tropopause:

\[
q_i(tpp) \approx 0 \quad q(tpp) \approx q_p
\]

However, diabatic effects are felt mainly through induced wind, \(u_i\)

\[
\frac{\partial q_p}{\partial t} + (u_p + \sum_i u_i) \nabla q_p \approx 0 \quad u_i = L^{-1}(q_i)
\]

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

\[
\frac{\bar{D}P}{Dt} = PV \cdot (\rho \mathbf{u}_j) + \nabla \cdot (\zeta \dot{\theta} + F \times \nabla \theta)
\]

Concentration/dilution of PV below/above heating
Non-advective transport of PV along \( \theta \) 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

\[ \theta = \text{constant} \]

Heating (no motion)

Motion (no heating)

\[ \rho P = \zeta \cdot \nabla \theta \]
Thought experiment

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

\[ \rho P = \zeta \cdot \nabla \theta \]
Thought experiment

\[ \theta = \text{constant} \]

Heating (no motion)  

Motion (no heating)

Now include some horizontal vorticity
Thought experiment

\[ \theta = \text{constant} \]

Heating (no motion) \rightarrow Motion (no heating)

Now include some horizontal vorticity

- **Vortex tilting** induced by response to heating \( \rightarrow \) 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:

\[
\frac{Du_g}{Dt} - fv_{ag} = 0
\]

\[
\frac{D\theta}{Dt} = H
\]

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

Define cross-frontal streamfunction, \( \psi \)

\[
v_{ag} = -\frac{\partial \psi}{\partial z} \quad w = \frac{\partial \psi}{\partial y}
\]

Meridional circulation acts to maintain TWB:

\[
\psi_{yy} \theta_z + \psi_{yz}(u_z - \theta_y) + \psi_{zz}(f - u_y) = H_y
\]
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\[+ w \frac{\partial}{\partial z} \]

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\[+ \psi_{zz} (f - u_v) = H_y\]
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time = 5.33 hours
Cls: \( \theta = 5.00 \) K, \( u = 5.00 \) m/s
min(PV) = -0.01 PVU, max(PV) = 2.92 PVU
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

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Thank you for your attention