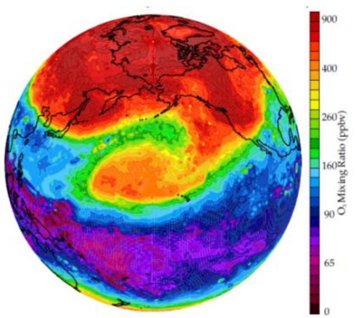
UTLS circulation and transport derived from satellite observations

- UTLS dynamics, circulation and transport
- Stratospheric temperature trends
- UTLS Asian monsoon
- Stratospheric water vapor
- Tropical tropopause layer (TTL)
- Tropical dynamics with GPS radio occultation data

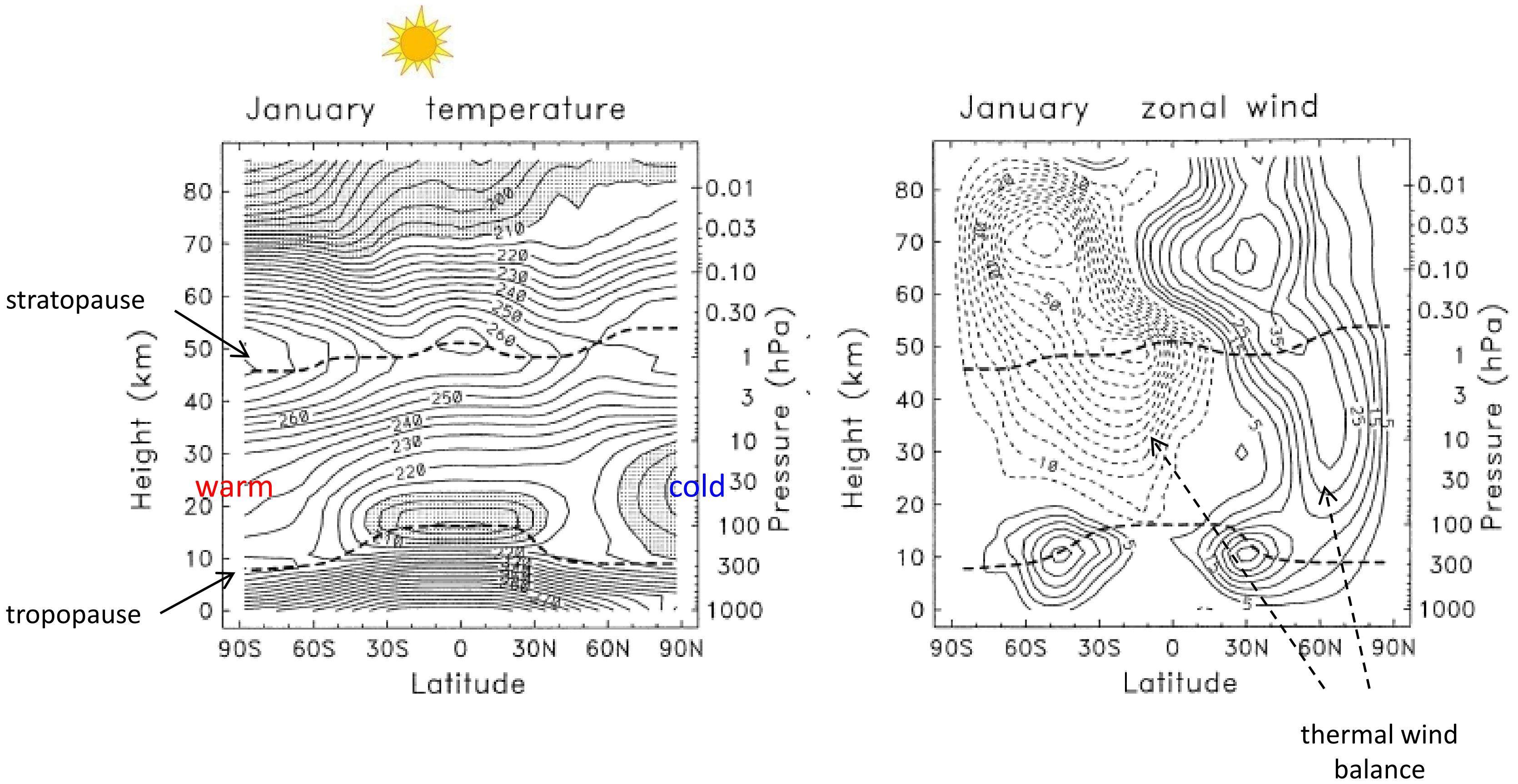
AIRS ozone 360 K



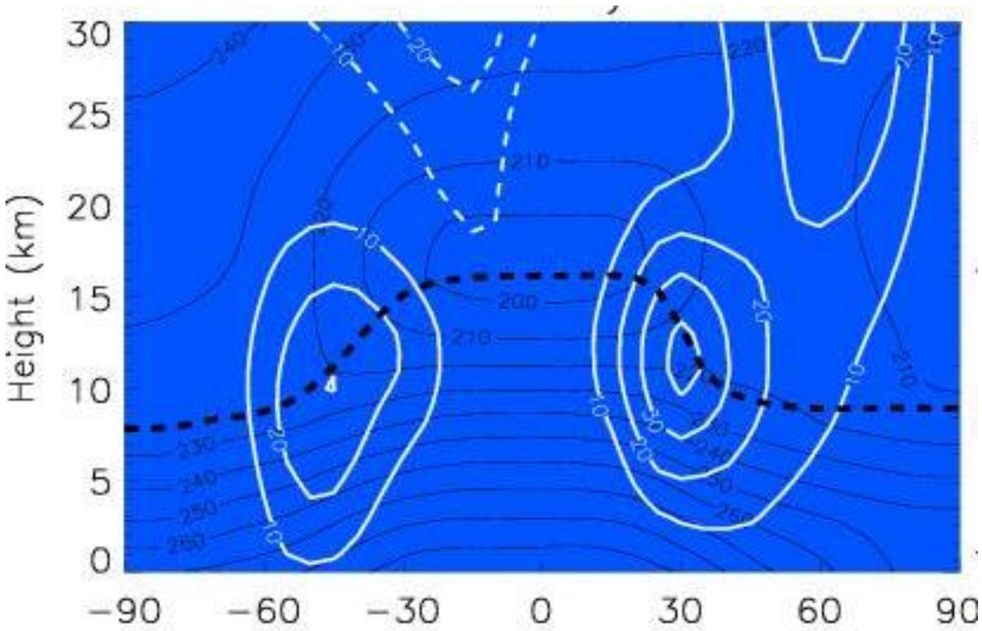
UTLS dynamics, circulation and transport

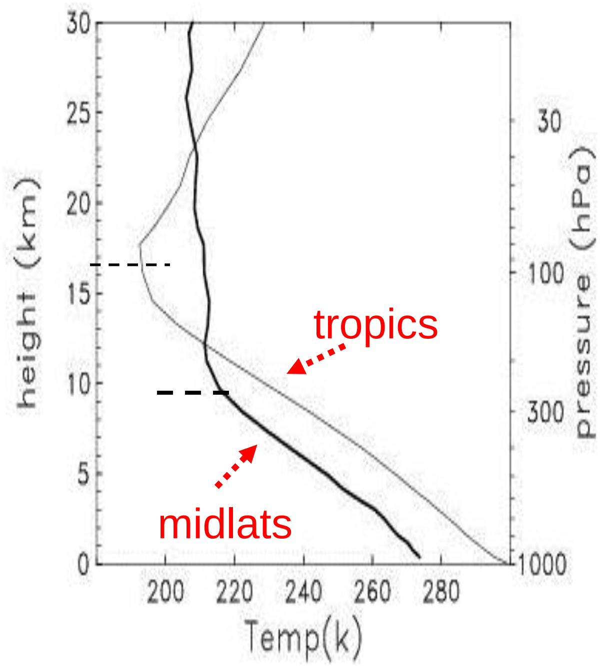
- Overview: why is the UTLS interesting?
- Circulation and variability of the stratosphere
- Rossby waves: mean flow forcing and dissipation
- Tropospheric baroclinic wave life cycles
- Large-scale tropical circulations
- Zonal mean constituent transport

Climatological temperatures and zonal winds in January



Global structure of the tropopause:





Strong change in stability across the tropopause:

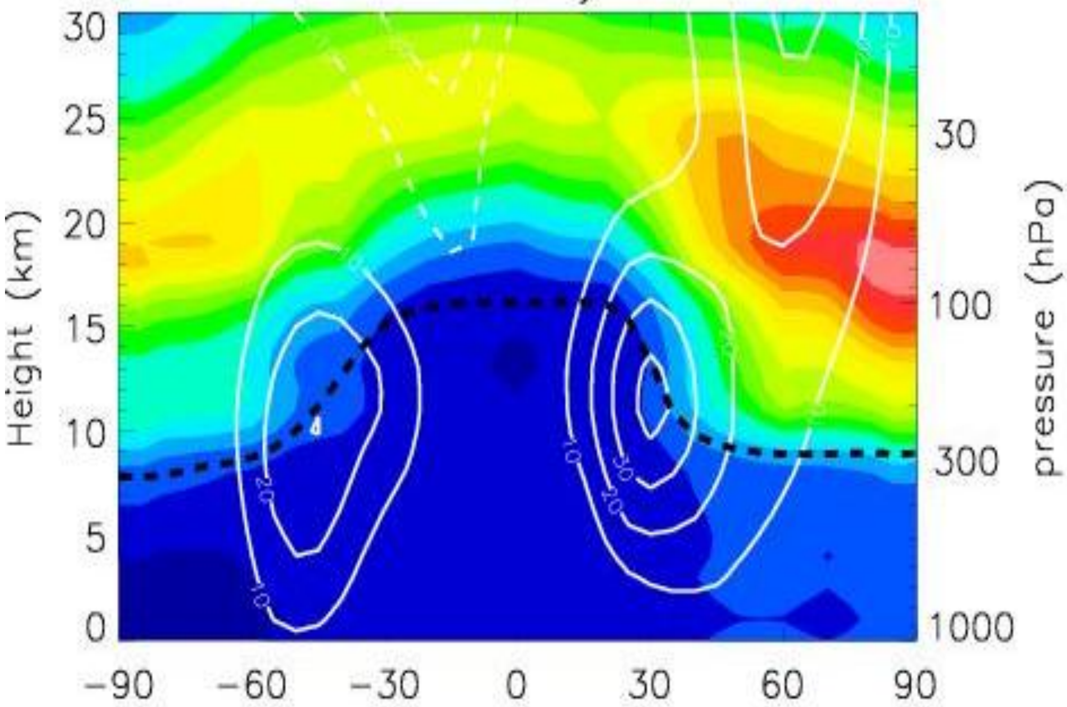
- Troposphere: vertically well-mixed; via convection and baroclinic instability
- Stratosphere: dynamically stable (mostly); circulation forced by radiation and forcing from troposphere (upward propagating waves)

Ozone

- Formed in stratosphere (stratospheric source gas)
- Long lifetime in lower stratophere
- Strong gradients across tropopause

Ozone column density, DU/km

Satellite climatology in January



Ozone

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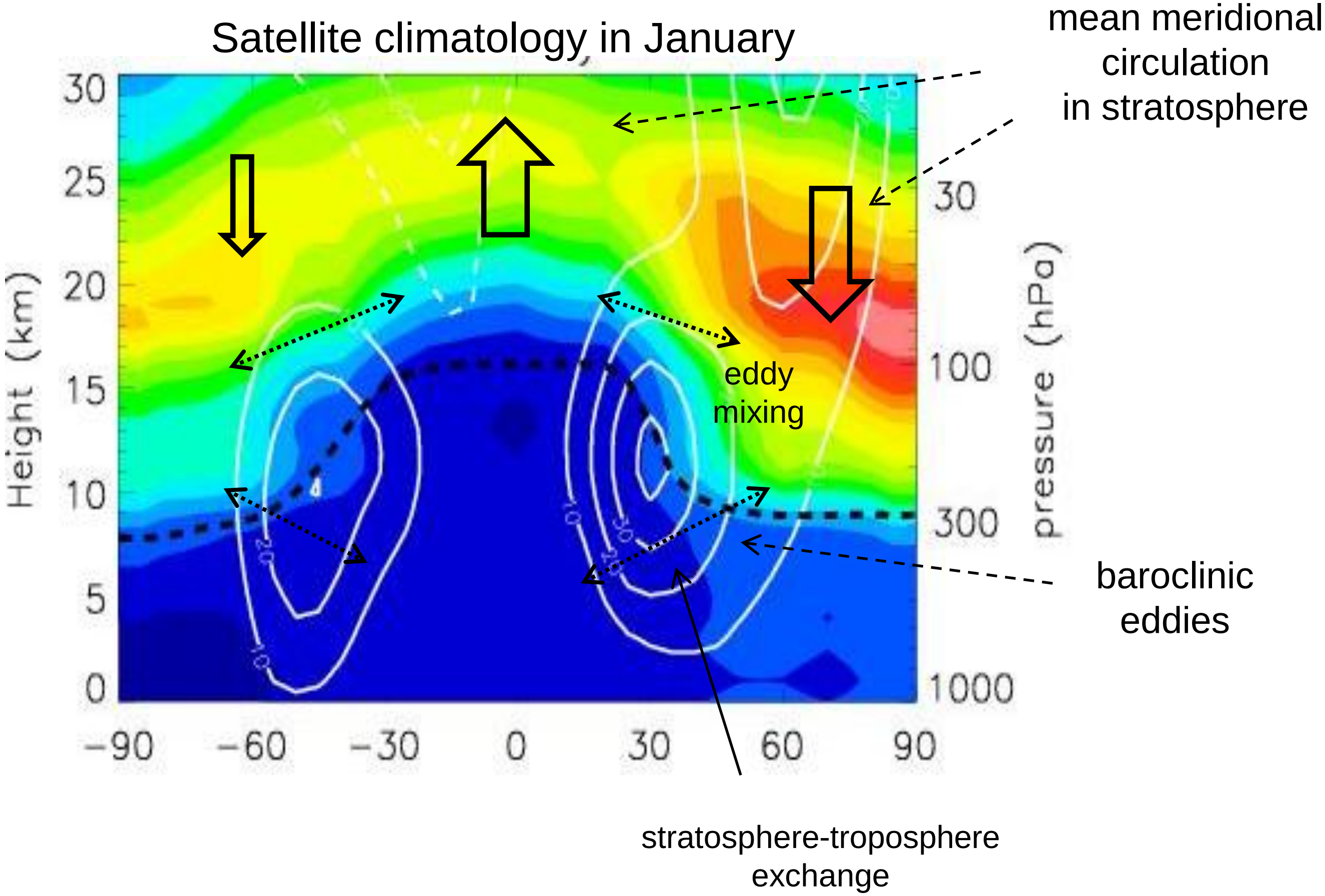
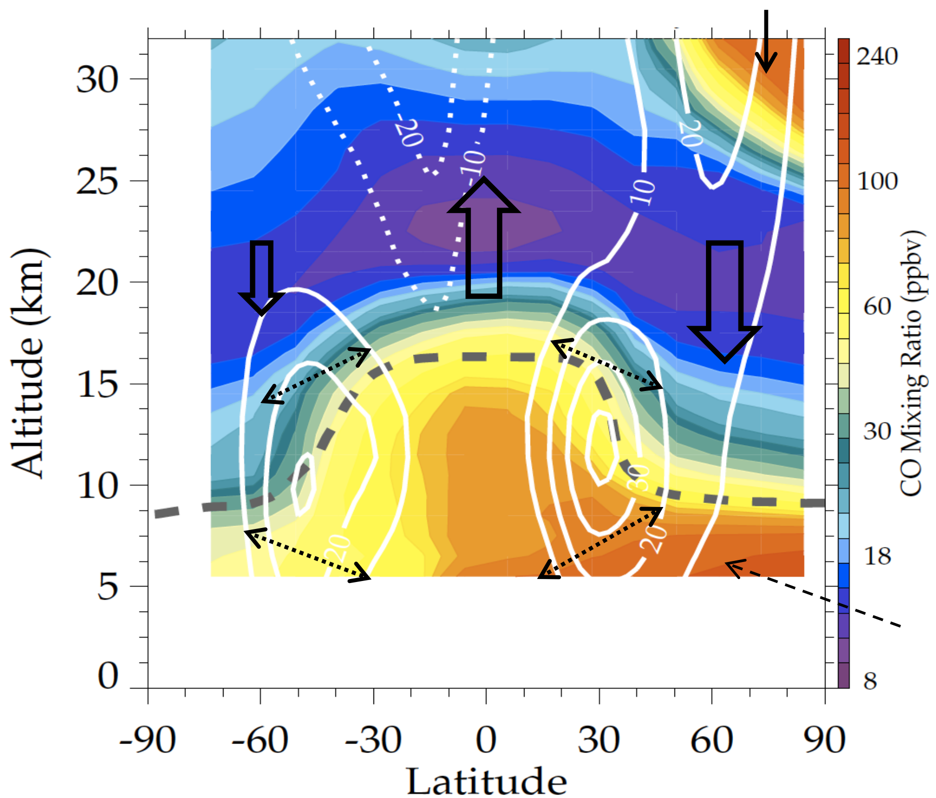
Ozone column density, DU/km



Carbon monoxide (CO)

- Emitted from combustion (tropospheric source gas)
- Photochemical lifetime of ~2 months (useful as a dynamical tracer)
- Strong gradients across tropopause

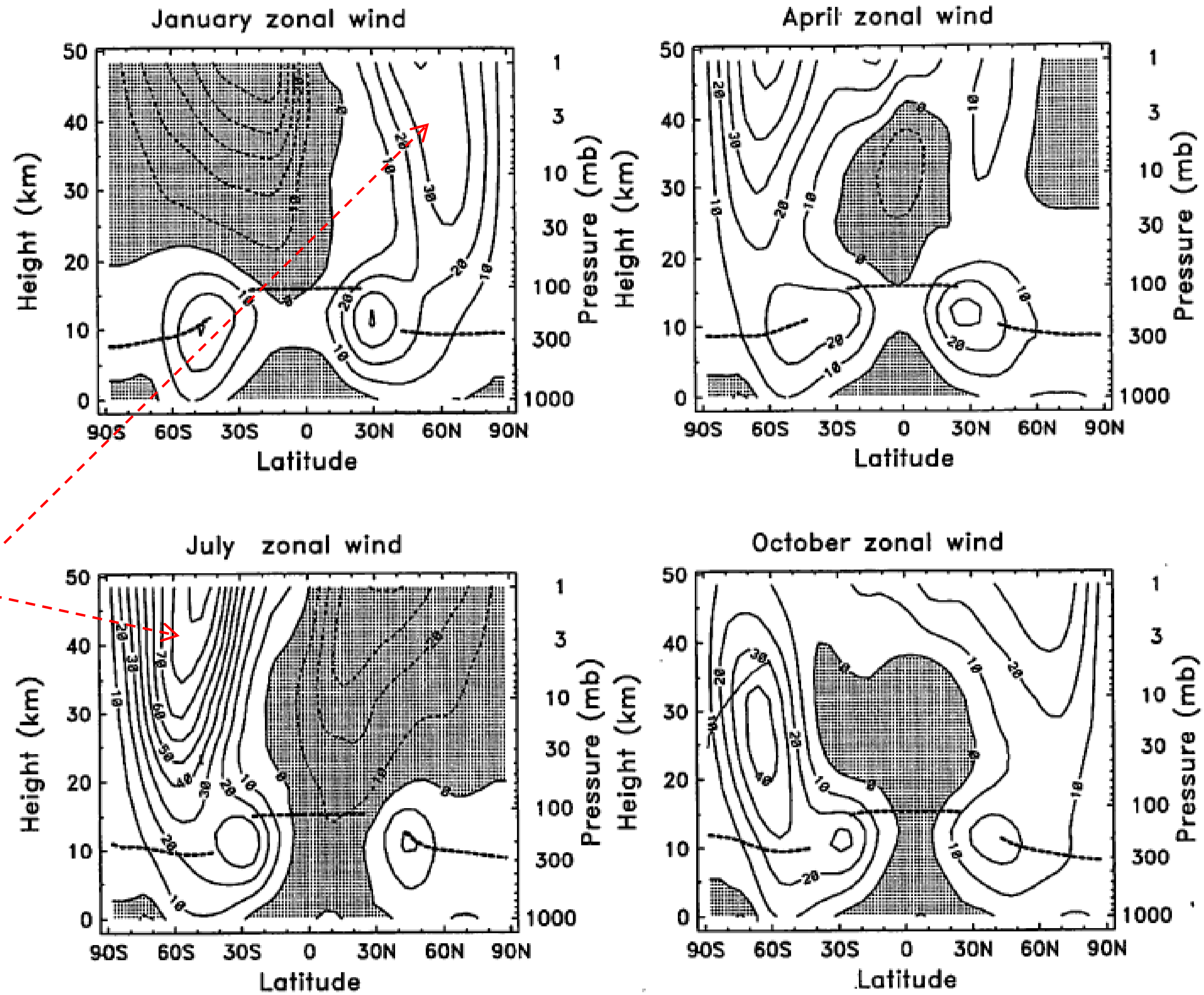
Measurements from ACE-FTS satellite



Strong gradients in chemical behavior demonstrates that the tropopause acts as a boundary separating distinct air masses

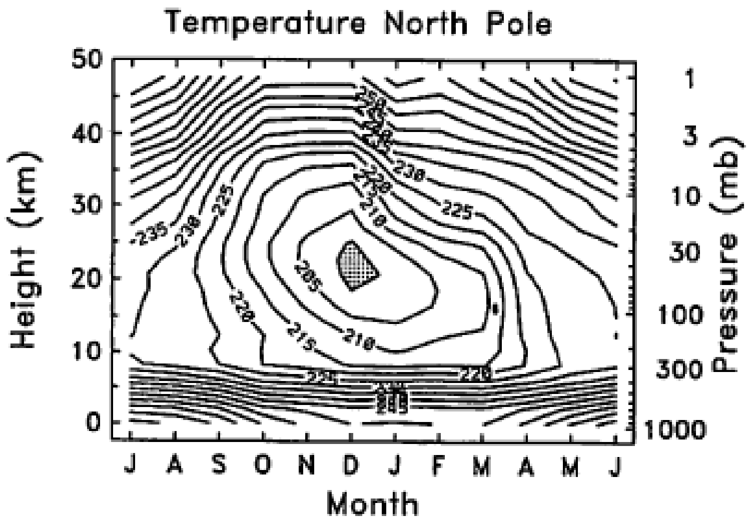
H_2O exhibits similar behavior

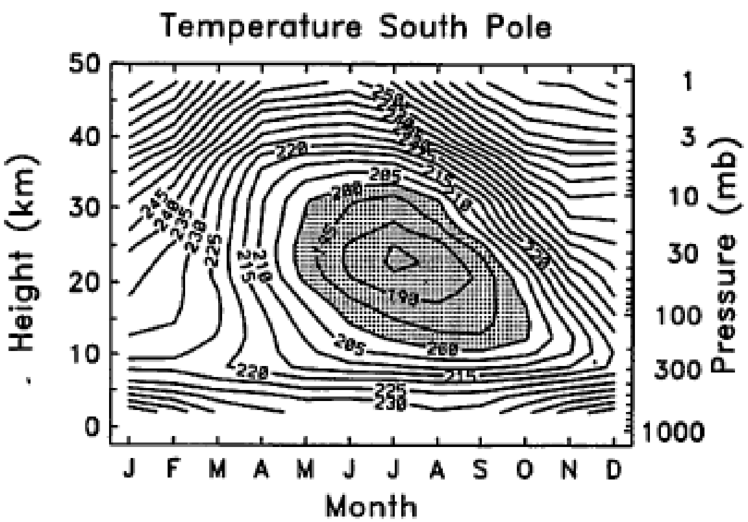
Stratospheric circulation: seasonal cycle of zonal mean winds



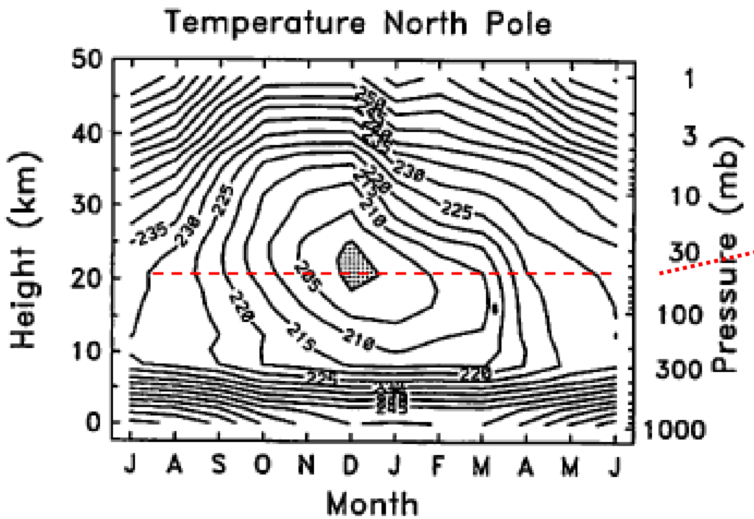
polar night jet
stronger in SH
than in NH

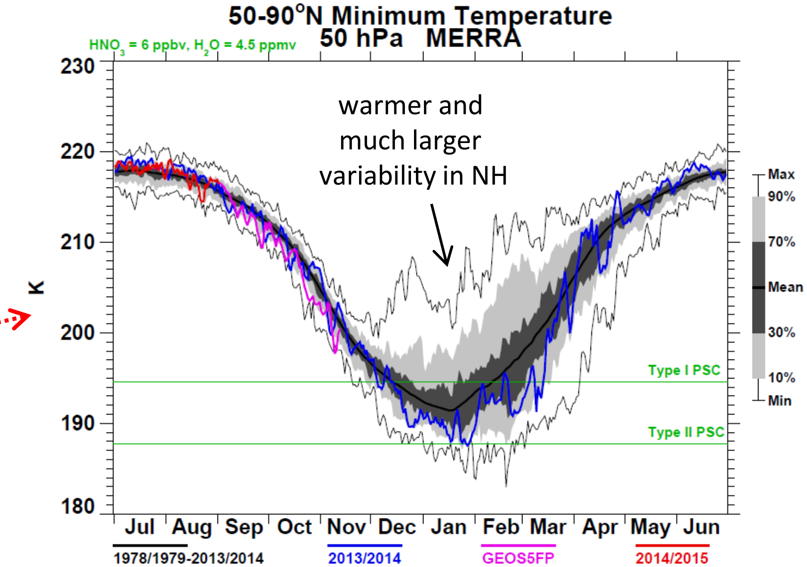
<u>climatological polar temperature</u>

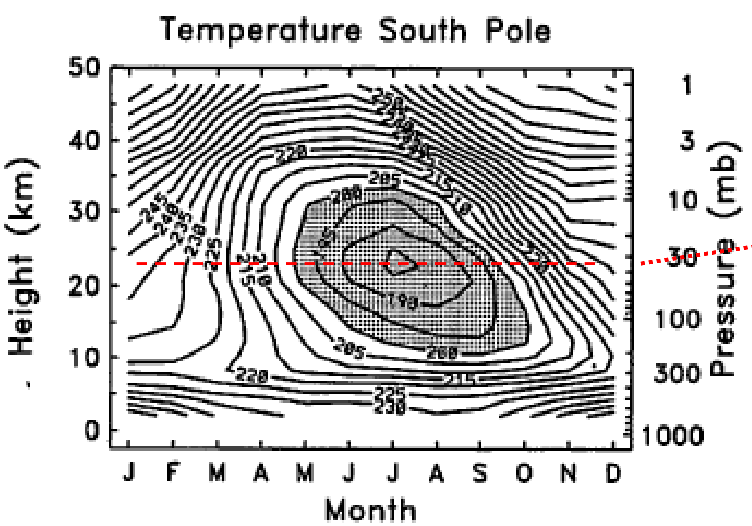


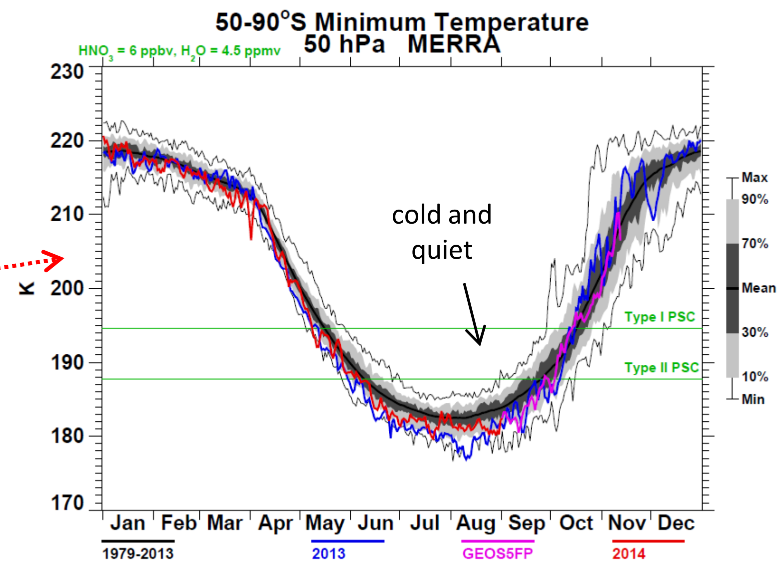


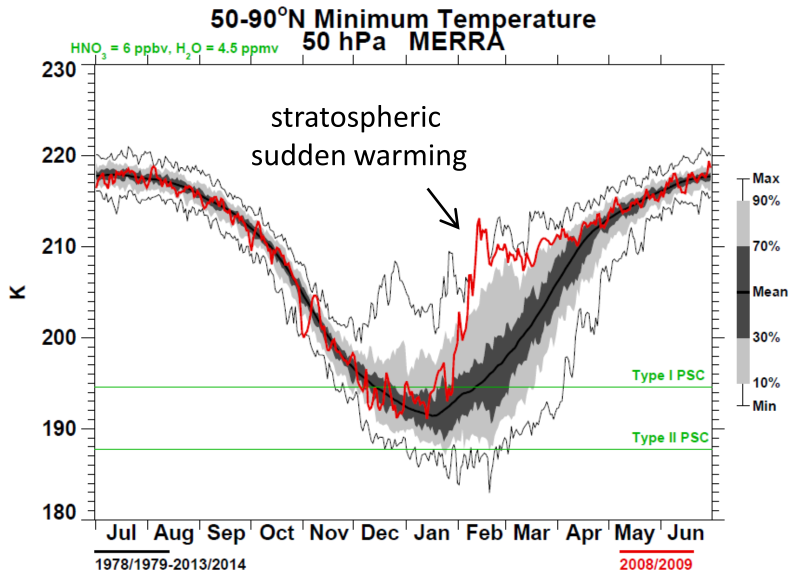
climatological polar temperature



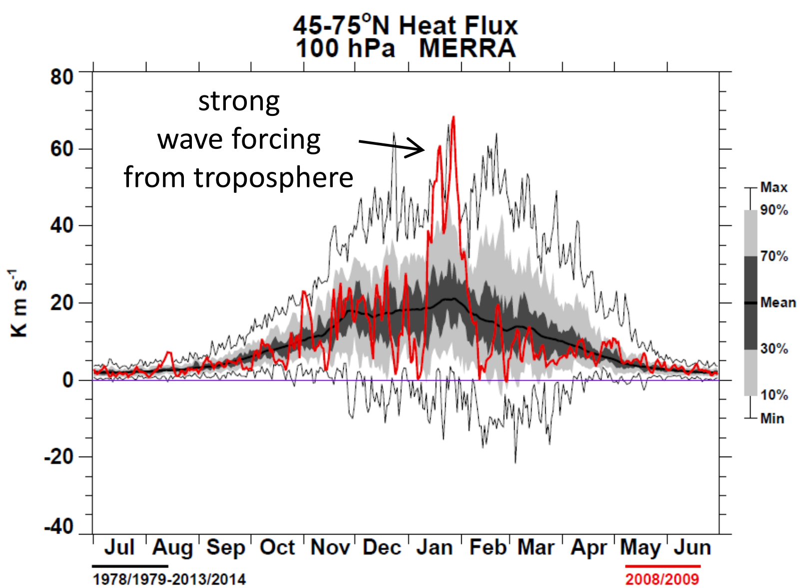


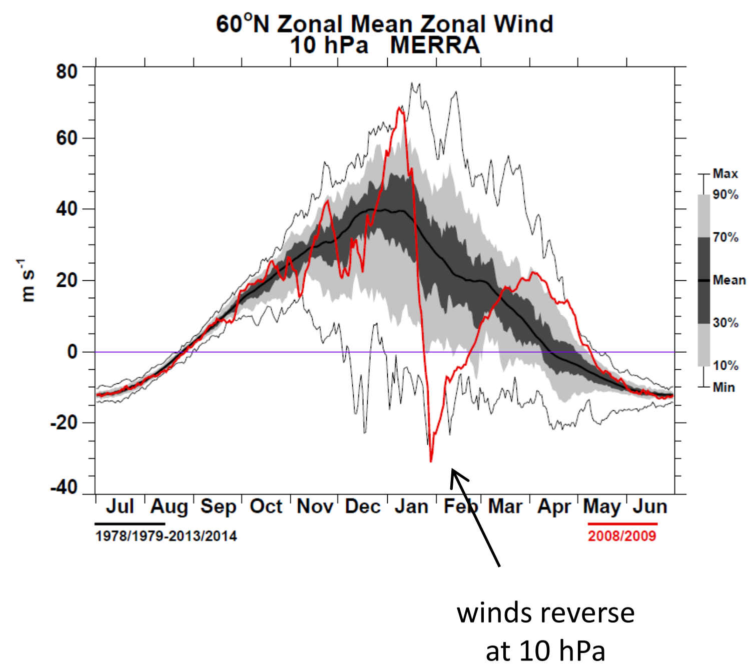






- Variability in NH winter stratosphere tied to large-scale forcing from troposphere.
- Episodic forcing produces 'stratospheric sudden warming' events.
- Largest observed stratosphere sudden warming in January 2009





A Major Stratospheric Sudden Warming Event in January 2009

Yayoi Harada, Atsushi Goto, Hiroshi Hasegawa, and Norihisa Fujikawa

Climate Prediction Division, Japan Meteorological Agency, Tokyo, Japan

JAS 2010

splitting of the polar vortex

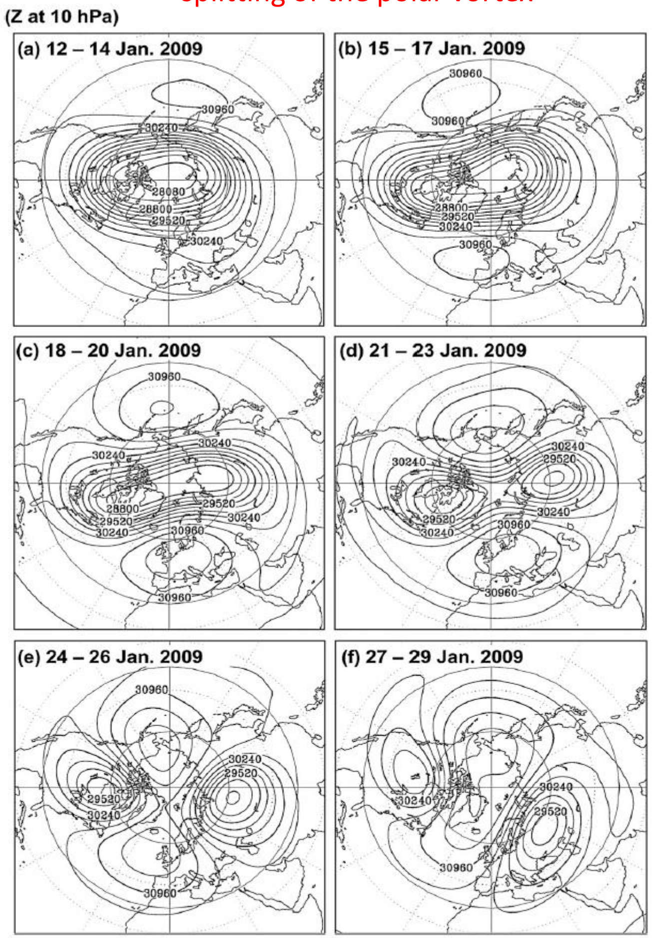
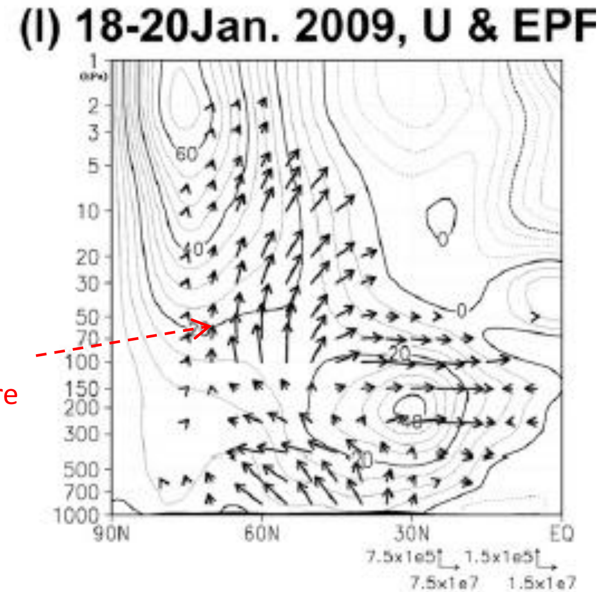


Fig. 3. The 10-hPa geopotential heights for six successive 3-day means in January 2009. The contour interval is 240 m.

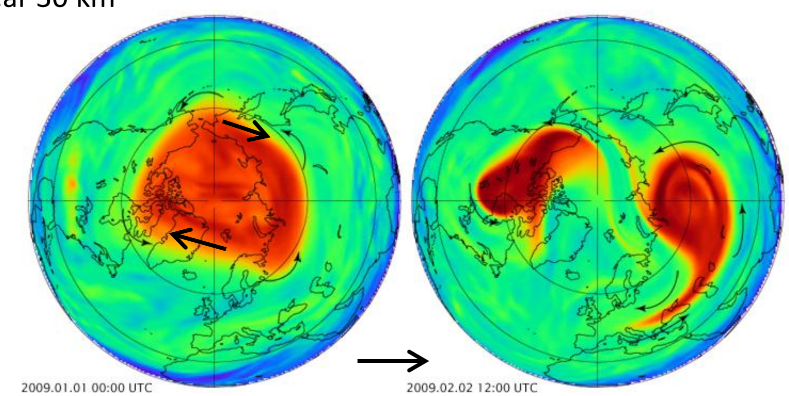
EP flux

wave forcing from troposphere



polar vortex near 30 km

potential vorticity



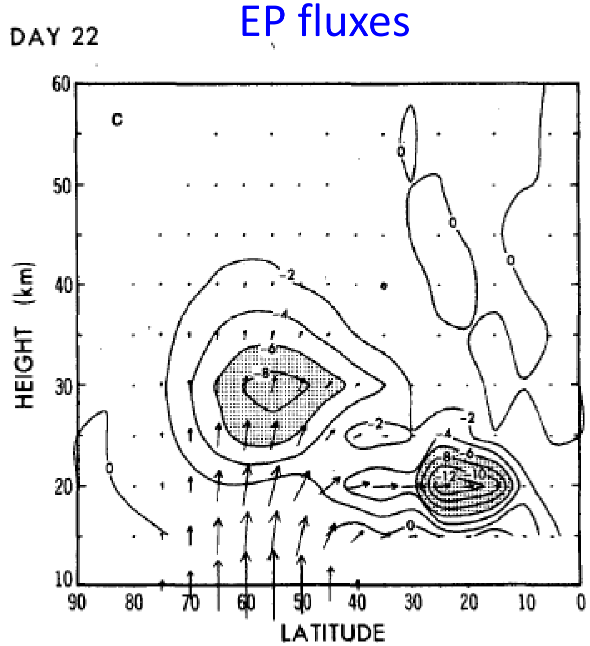
split of polar vortex

A Dynamical Model of the Stratospheric Sudden Warming

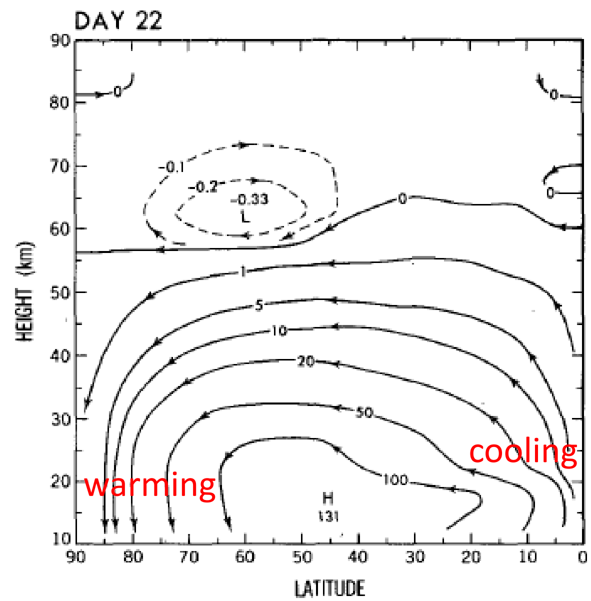
Taroh Matsuno[1]
Geophysical Fluid Dynamics Laboratory, NOAA, Princeton University, Princeton, N. J.
(Manuscript received 29 March 1971, in revised form 16 August 1971)

solution to puzzle of stratospheric warmings

EP fluxes



overturning circulation



Some Eulerian and Lagrangian Diagnostics for a Model Stratospheric Warming[1]

T. Dunkerton, C.-P. F. Hsu[2] and M. E. McIntyre[3]
Department of Atmospheric Sciences, University of Washington, Seattle 98195
(Manuscript received 30 May 1980, in final form 11 December 1980)

JAS 1981

Governing equations for the zonal mean flow (Transformed Eulerian mean)

EP flux divergence (wave forcing)

zonal momentum balance

$$\frac{\partial \overline{u}}{\partial t} - \hat{f}\overline{v}^{*} = \mathrm{DF}$$

diabatic forcing

thermodynamic balance

$$\frac{\partial \overline{T}}{\partial t} + \overline{v}^{*}\frac{1}{a}\frac{\partial \overline{T}}{\partial \phi} + \overline{w}^{*}S = \overline{Q},$$

continuity equation

$$(a\cos\phi)^{-1}\frac{\partial}{\partial \phi}(\overline{v}^{*}\cos\phi) + e^{z/H}\frac{\partial}{\partial z}(\overline{w}^{*}e^{-z/H}) = 0,$$

geostrophic thermal wind

$$f\frac{\partial \overline{u}}{\partial z} + \frac{R}{aH}\frac{\partial \overline{T}}{\partial \phi} = 0.$$

Andrews et al, 1987

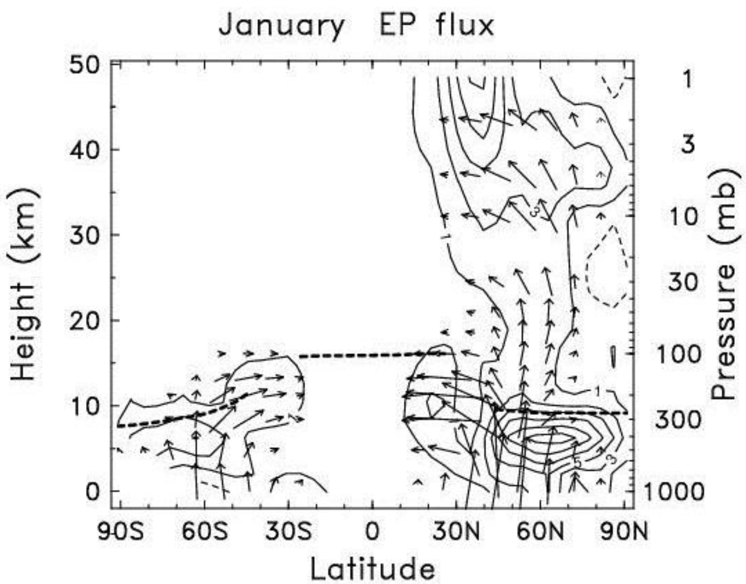
Eliassen-Palm fluxes:

EP flux divergence (wave forcing)

$$\frac{\partial \overline{u}}{\partial t} - \hat{f}\overline{v}^{*} = \mathrm{DF}$$

$$\mathrm{DF} = \frac{\exp(z/H)}{a\ \cos\phi} \nabla \cdot \mathbf{F},$$

climatology



components:

momentum flux

latitudinal flux

$$F_{\phi} = \exp(-z/H) a\ \cos\phi \left[-\overline{u'v'} \right],$$

vertical flux

$$F_{z} = \exp(-z/H) a\ \cos\phi \left[\hat{f}\frac{\overline{v'T'}}{S} \right],$$

heat flux

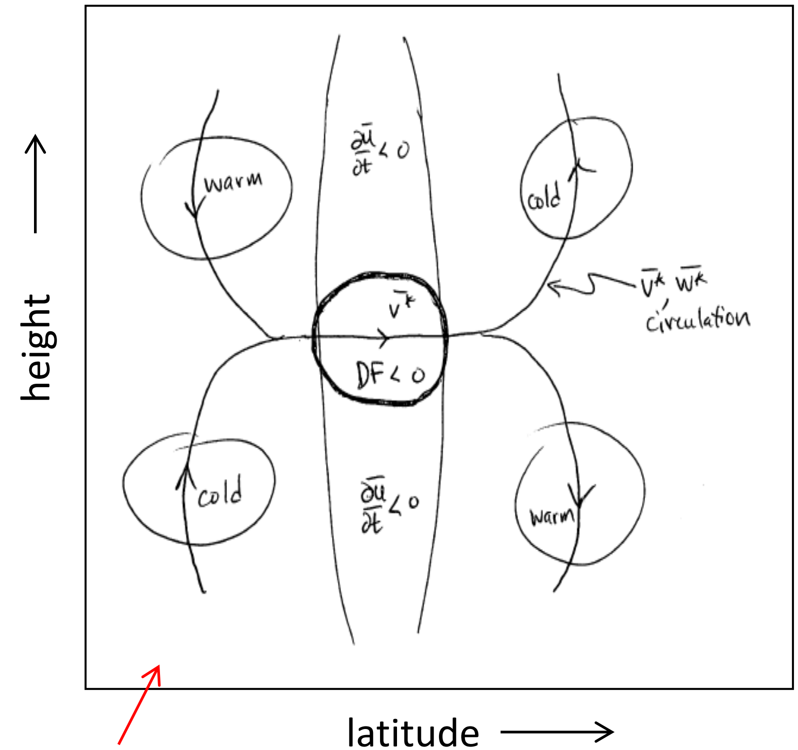
Important points:

- DF quantifies zonal momentum forcing
- **F** proportional to ‘wave activity’ flux
  (DF shows sources and sinks of waves)
- F_{ϕ} and F_z indicate direction of wave propagation

Response of a balanced vortex to localized EP flux forcing (DF)

$$\frac{\partial \bar{u}}{\partial t} - f\,\overline{v}^* = DF$$

- response is balanced between $\partial\bar{u}/\partial t$ and $f\,\overline{v}^*$
- $\overline{v}^*$, $\overline{w}^*$ and $\partial\bar{u}/\partial t$ act to extend DF forcing non-locally
- overall circulation maintains thermal wind balance



effects of forcing extended in altitude and latitude

Circulation response depends on frequency of forcing:

Combine equations:

time dependence

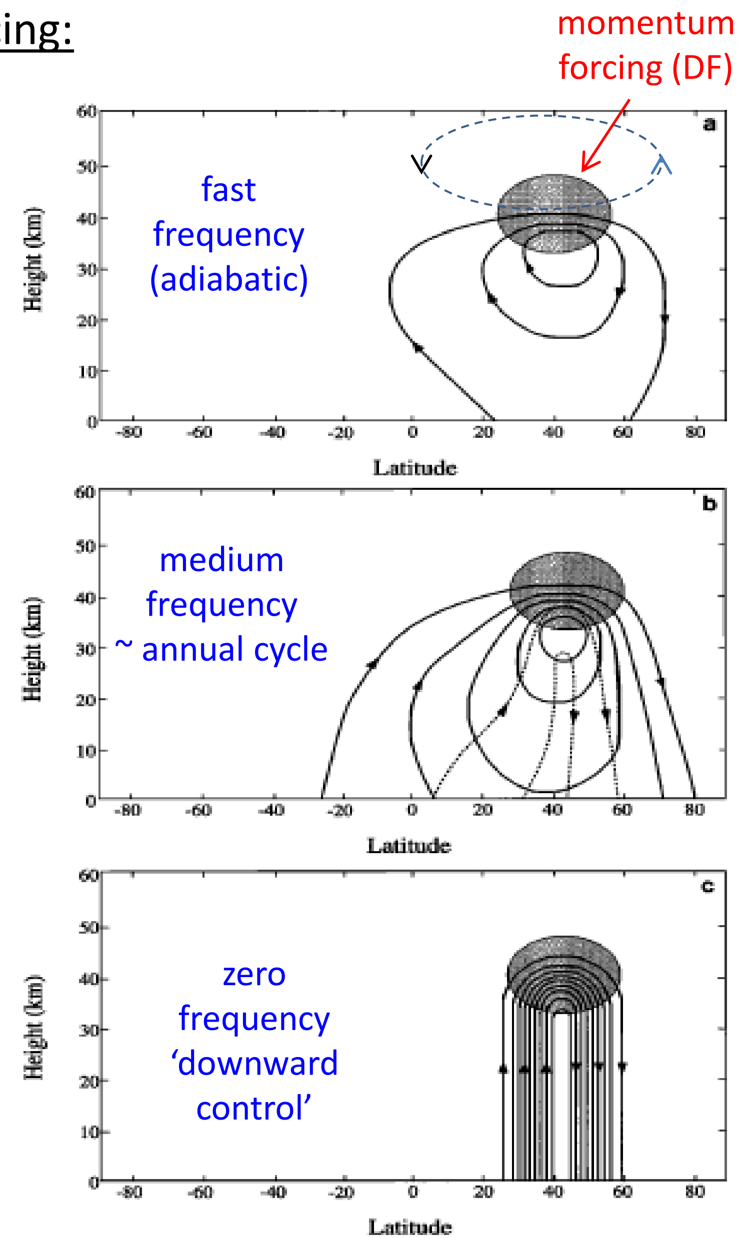
$$\frac{\partial}{\partial z}\left(\frac{1}{\rho_0}\frac{\partial(\rho_0\hat{w})}{\partial z}\right) + \left(\frac{i\sigma}{i\sigma+\alpha}\right)\frac{N^2}{4\Omega^2a^2\cos\phi}\frac{\partial}{\partial\phi}\left(\frac{\cos\phi}{\sin^2\phi}\frac{\partial\hat{w}}{\partial\phi}\right)$$

$$= \left(\frac{i\sigma}{i\sigma+\alpha}\right)\frac{(R/H)}{4\Omega^2a^2\cos\phi}\frac{\partial}{\partial\phi}\left(\frac{\cos\phi}{\sin^2\phi}\frac{\partial\hat{Q}}{\partial\phi}\right)$$

diabatic heating

$$+ \frac{1}{2\Omega a\cos\phi}\frac{\partial}{\partial\phi}\left(\frac{\cos\phi}{\sin\phi}\frac{\partial\hat{G}}{\partial z}\right)$$

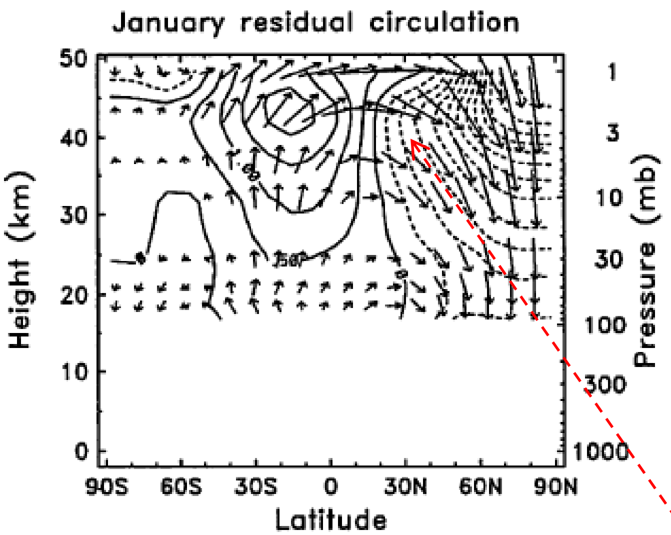
momentum forcing (DF)

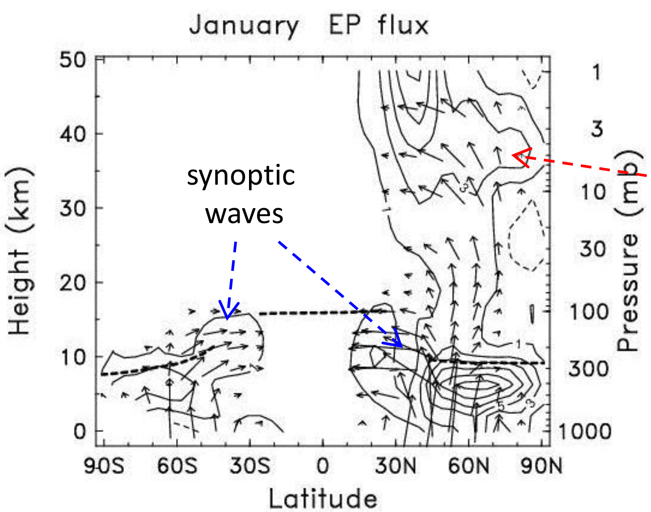


In general both Q and DF drive the mean circulation. These plots show the response to isolated forcing from Rossby wave EP flux divergence. The lower cell becomes more important for slower forcing.

Haynes et al 1991
Holton et al 1995

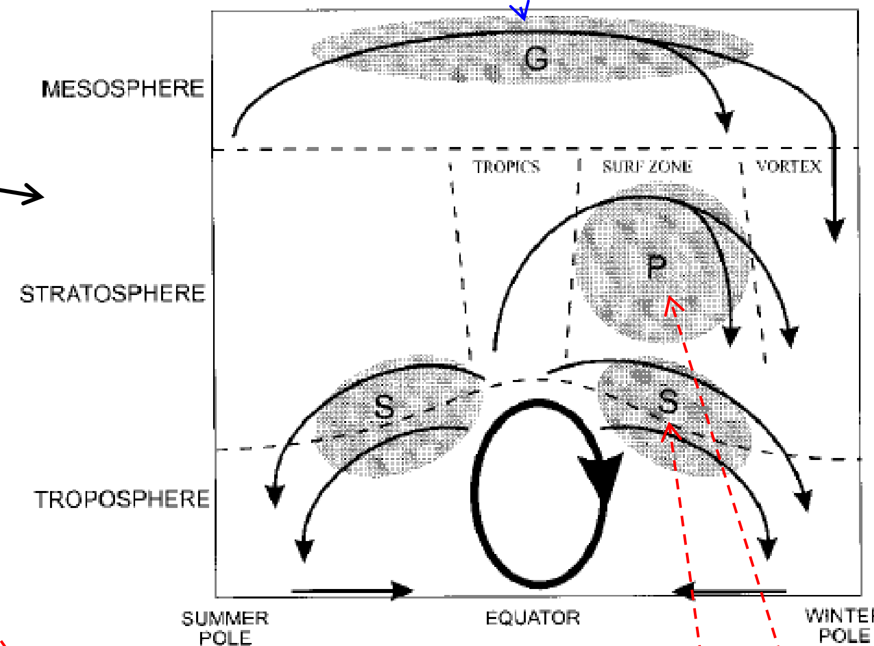
Climatology of stratospheric overturning circulation





gravity waves

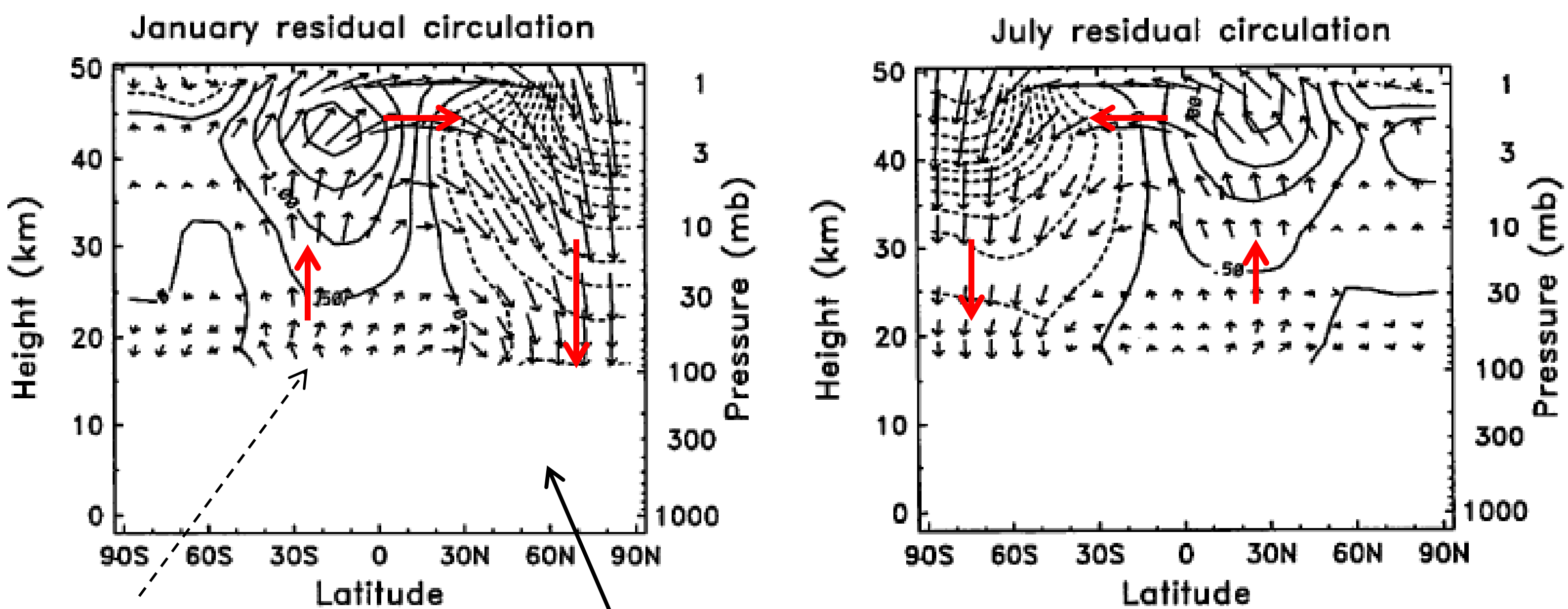
Plumb 2002



winter planetary waves help drive the overturning circulation

upper and lower branches of the circulation driven by **Planetary** and **Synoptic** waves

The overturning circulation reverses between solstice seasons



stronger tropical upwelling during boreal winter

circulation is stronger during NH winter, related to stronger wave forcing from troposphere

The stratospheric overturning circulation is often termed the Brewer-Dobson circulation (closely related to the Lagrangian or transport circulation)

deduced by Brewer (1949) studying stratospheric water vapor and Dobson (1956) studying stratospheric ozone

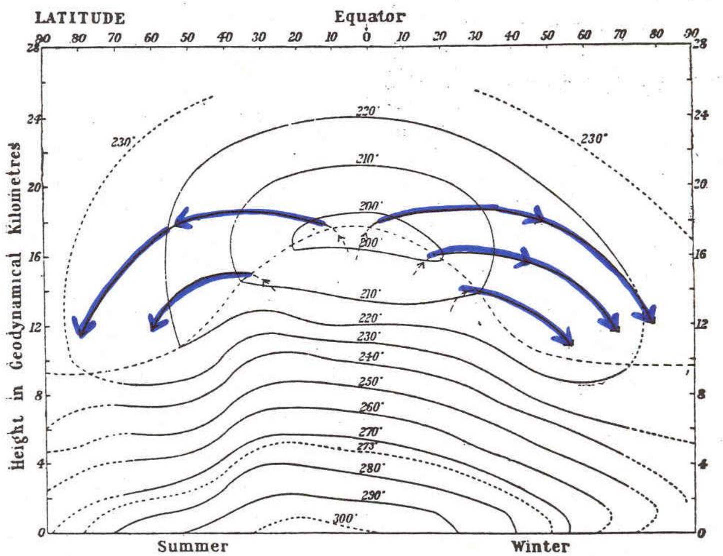
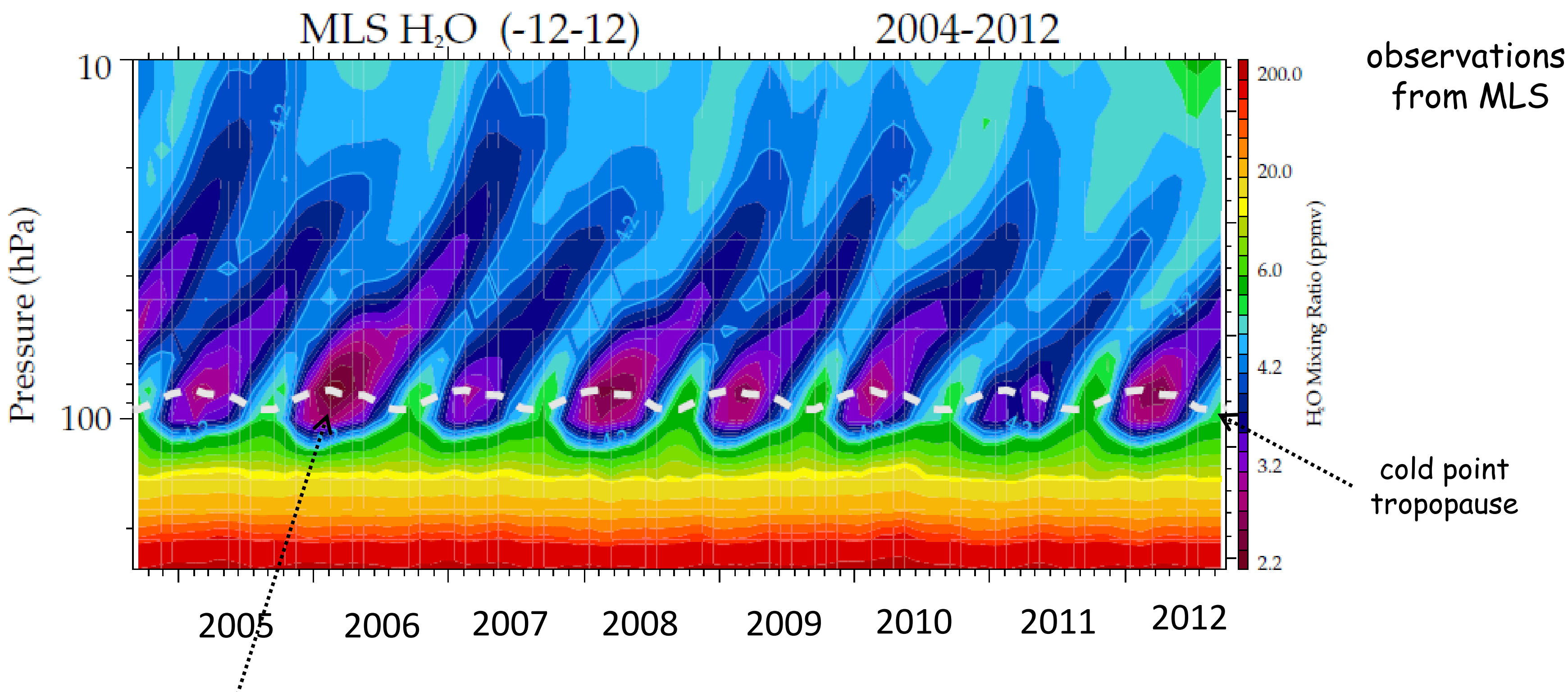


FIG. 5. A supply of dry air is maintained by a slow mean circulation from the equatorial tropopause.

see recent review by Butchart 2014

Annual cycle: stronger tropical upwelling, colder temperatures and low H_2O



- annual cycle in tropopause temperature imparts annual cycle in H_2O
- upward propagation with Brewer-Dobson circulation

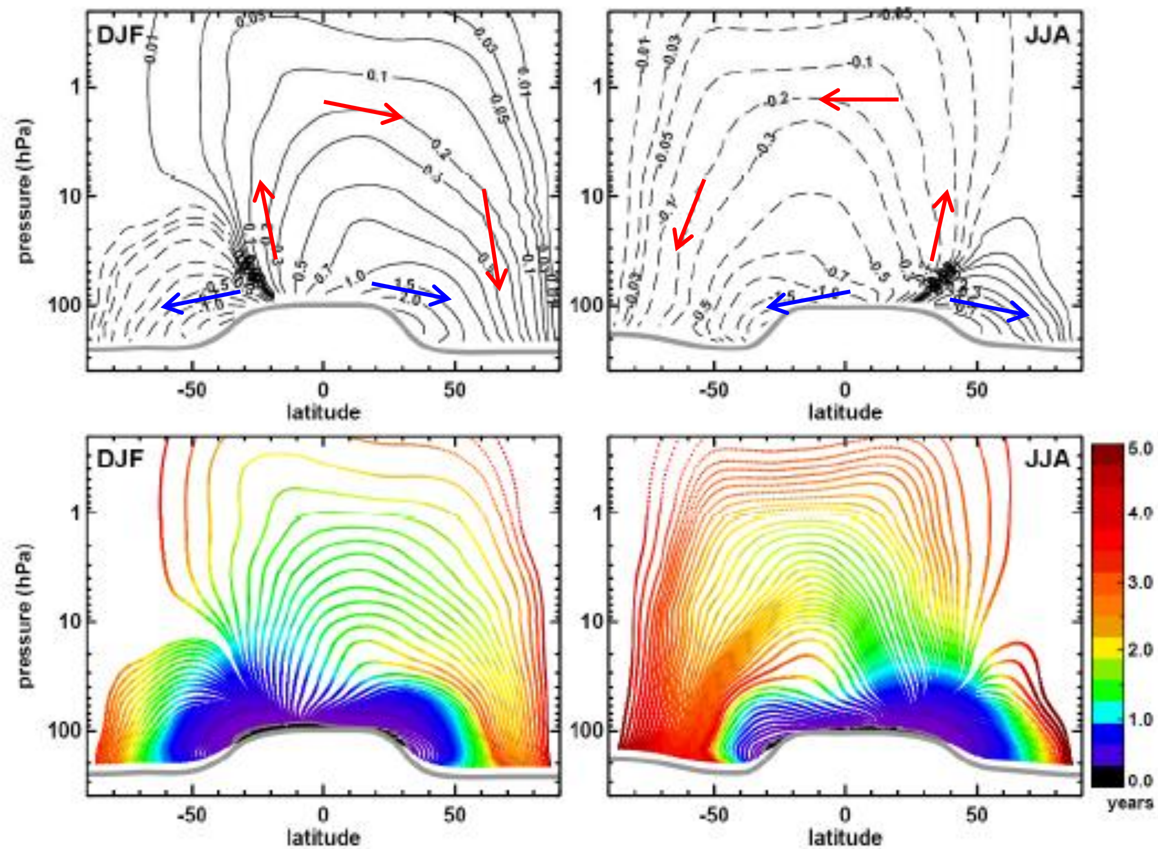
Residual circulation trajectories and transit times into the extratropical lowermost stratosphere

T. Birner[1] and H. Bönisch[2]

ACP 2011

renewed appreciation that there are upper and lower branches of the BDC

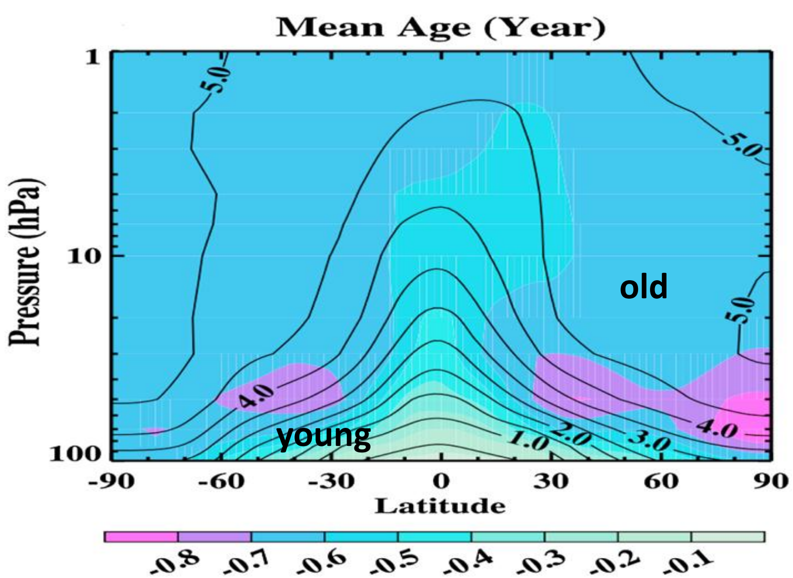
Red: deep branch (slow)
Blue: shallow branch (fast)



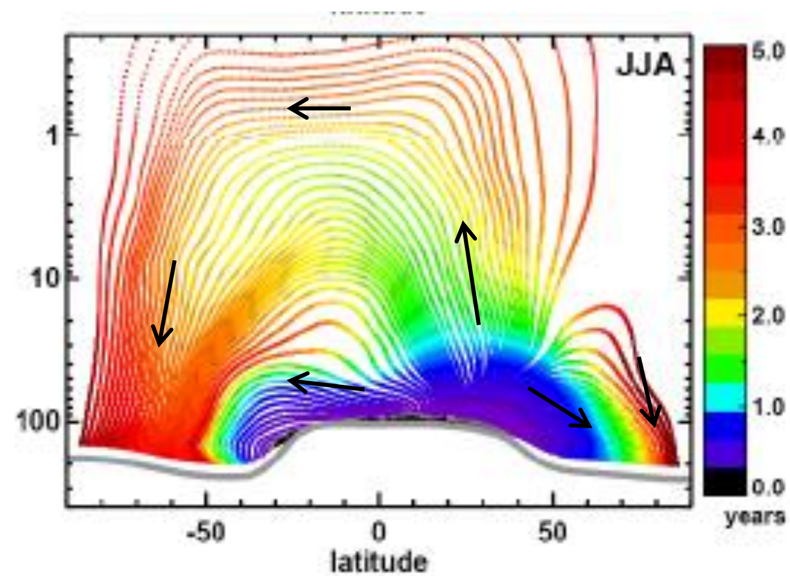
streamfunction

transit time along constant streamlines

Transit time is closely related to 'mean age' (time since air entered stratosphere)



transit time



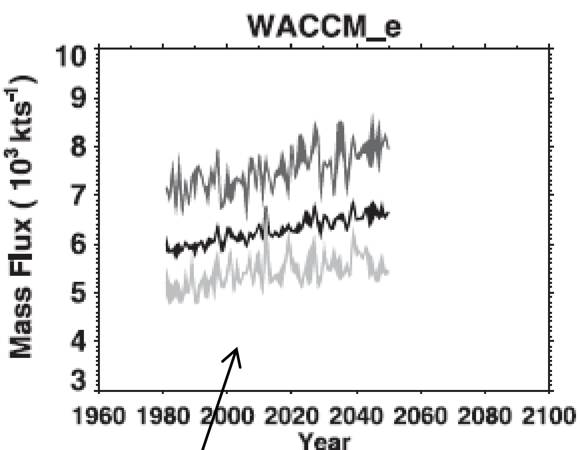
Air at any particular location is characterized by a distribution of transit times and ages (so-called age spectrum)

Chemistry–Climate Model Simulations of Twenty-First Century Stratospheric Climate and Circulation Changes

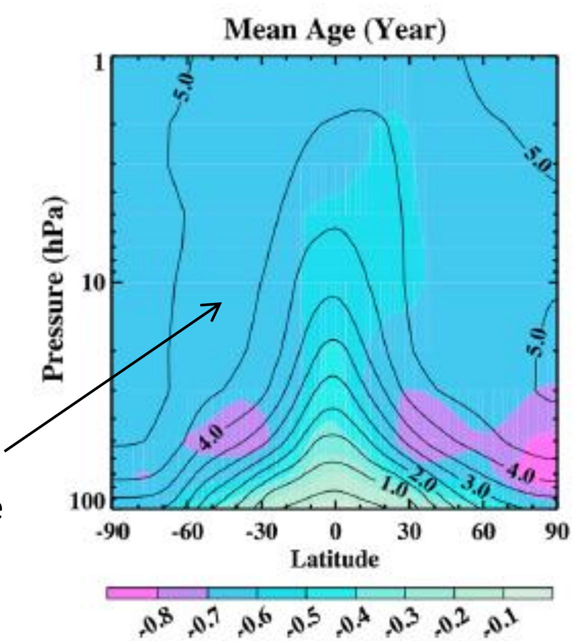
NEAL BUTCHART,[a] I. CIONNI,[b] V. EYRING,[b] T. G. SHEPHERD,[c] D. W. WAUGH,[d] H. AKIYOSHI,[e] J. AUSTIN,[f] C. BRÜHL,[g] M. P. CHIPPERFIELD,[h] E. CORDERO,[i] M. DAMERIS,[b] R. DECKERT,[b] S. DHOMSE,[h] S. M. FRITH,[j] R. R. GARCIA,[k] A. GETTELMAN,[k] M. A. GIORGETTA,[l] D. E. KINNISON,[k] F. LI,[m] E. MANCINI,[n] C. MCLANDRESS,[c] S. PAWSON,[o] G. PITARI,[n] D. A. PLUMMER,[p] E. ROZANOV,[q] F. SASSI,[r] J. F. SCINOCCA,[s] K. SHIBATA,[t] B. STEIL,[g] AND W. TIAN[h]

JGR 2010

increasing BDC circulation in models

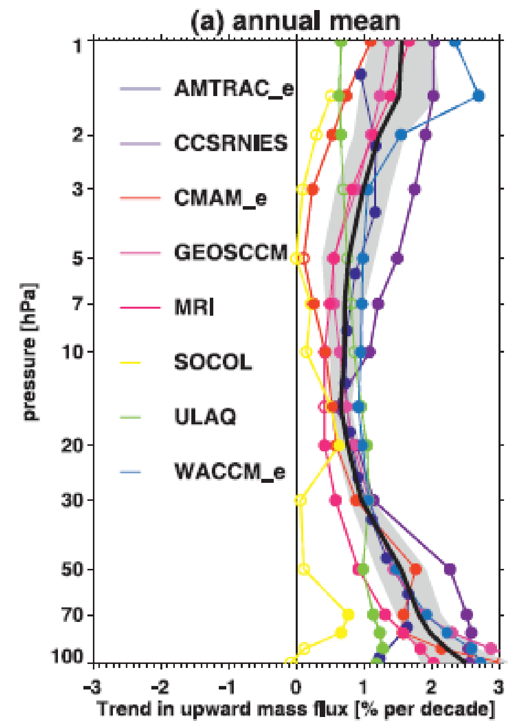


increasing tropical upwelling



younger age

Colors: age trends



faster upwelling

Key points:

- Asymmetry in winter stratosphere circulations: more disturbed in the NH, cold and quiet in the SH

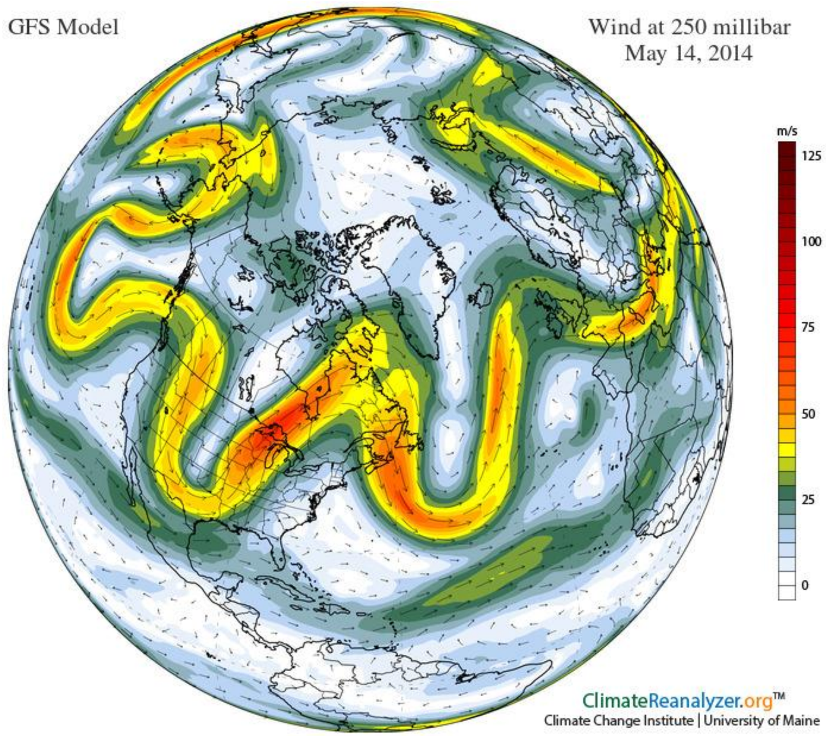
- Stratospheric circulation is forced by waves from the troposphere (stronger forcing in NH; episodic stratospheric sudden warmings)

- Dynamical response of balanced vortex to wave forcing (non-local temperature and wind changes)

- Eliassen-Palm (EP) fluxes quantify wave forcing

- Brewer-Dobson transport circulation (deep and shallow branches)

Rossby waves



Rossby wave propagation: quasi-geostrophic linearized PV equation

$$\left(\frac{\partial}{\partial t}+\frac{\bar{u}}{a\cos\phi}\frac{\partial}{\partial\lambda}\right)q'_{(\mathrm{M})}+a^{-1}\bar{q}_\phi v'=0,$$

eddy PV

background PV gradient

$$\bar{q}_\phi = 2\Omega\cos\phi - \left[\frac{(\bar{u}\cos\phi)_\phi}{a\cos\phi}\right]_\phi - \frac{a}{\rho_0}\left(\frac{\rho_0 f^2}{N^2}\bar{u}_z\right)_z.$$

wave solution:

$$\Phi' = e^{z/2H}\,\mathrm{Re}\,\Psi(\phi, z)e^{is\lambda}$$

$$\boxed{\frac{f^2}{a^2\cos\phi}\left(\frac{\cos\phi}{f^2}\Psi_\phi\right)_\phi+\frac{f^2}{N^2}\Psi_{zz}+n_s^2\Psi=0}$$

wave equation:
propagation for
$n_s^2 > 0$

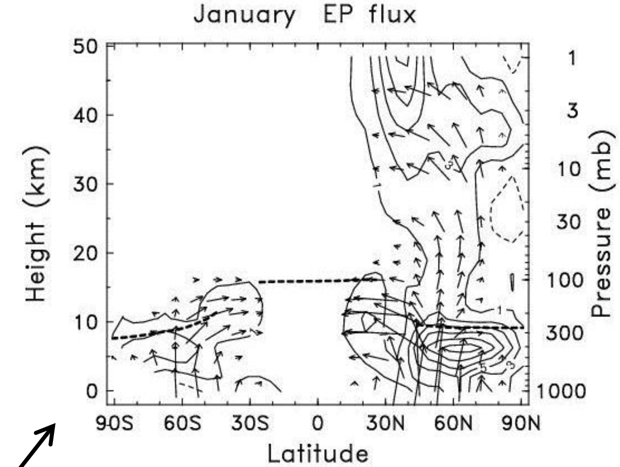
$$n_s^2=\frac{\bar{q}_\phi}{a\bar{u}}-\frac{s^2}{a^2\cos^2\phi}-\frac{f^2}{4N^2H^2}$$

refractive index

Propagation of Planetary-Scale Disturbances from the Lower into the Upper Atmosphere

J. G. Charney and P. G. Drazin[1]

JGR 1961

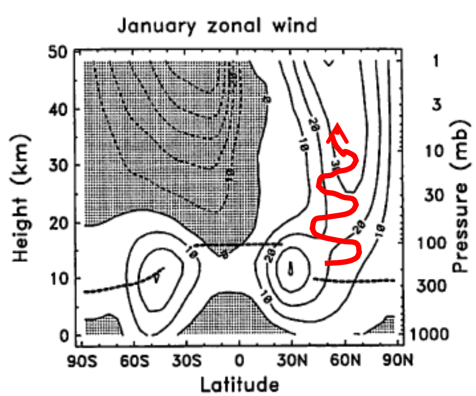


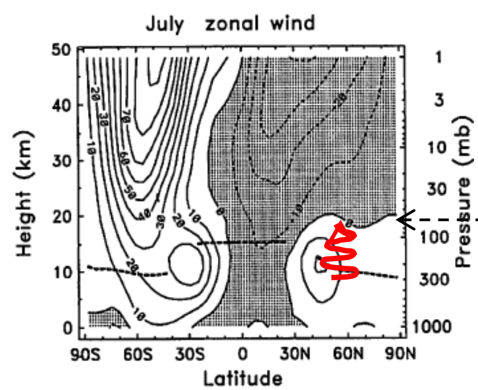
~ cos (lat) + U terms

$$n_s^2 = \frac{\bar{q}_\phi}{a\bar{u}} - \frac{s^2}{a^2\cos^2\phi} - \frac{f^2}{4N^2H^2}$$

<u>2 key points:</u>

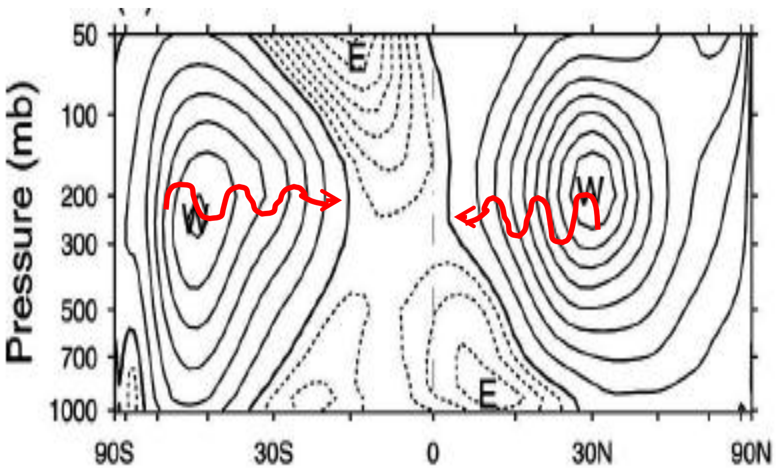
- n_s^2 proportional to ~cos (lat) (Rossby wave refraction towards low latitudes)
- vertical propagation for U > 0 and small zonal wavenumbers
  (planetary waves propagate to stratosphere only during winter)

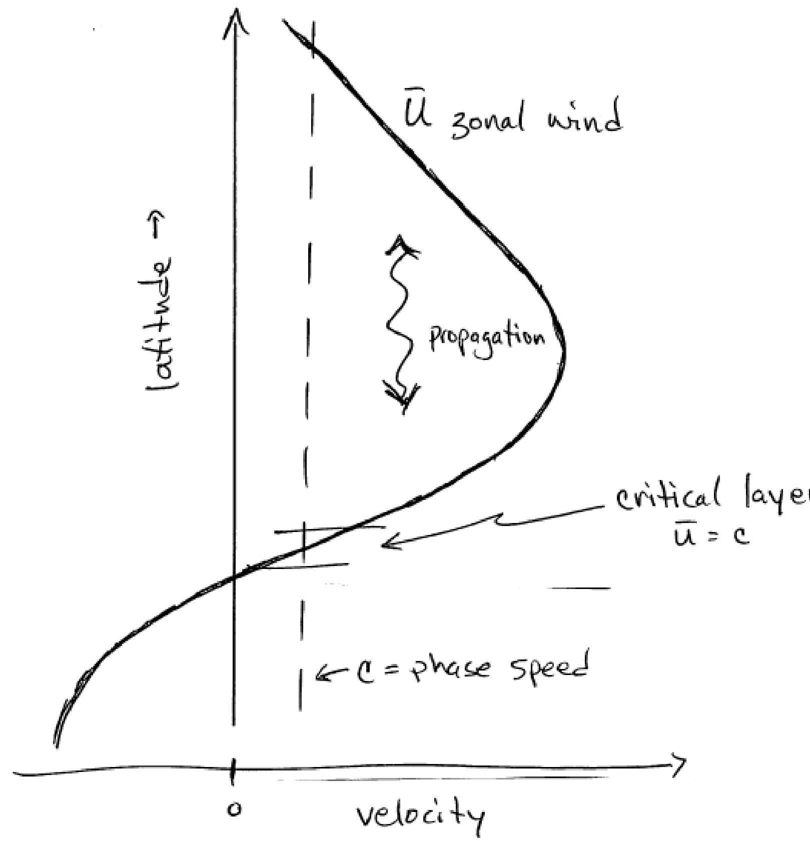




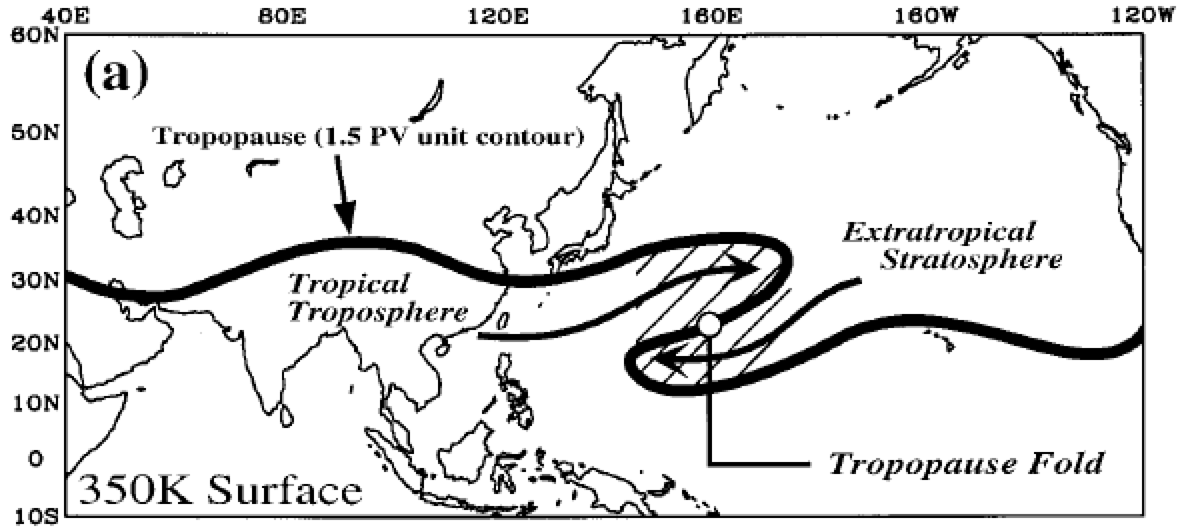
no propagation in summer easterlies

Rossby waves cannot propagate into regions where U < 0   (i.e. across equator)



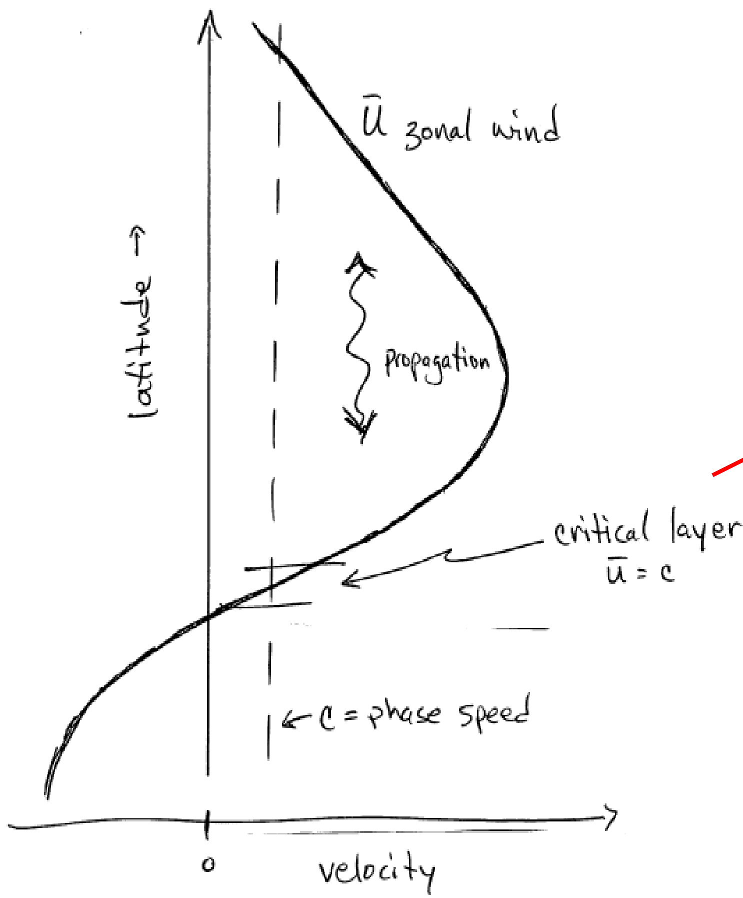


Breaking Rossby waves: overturning of PV contours

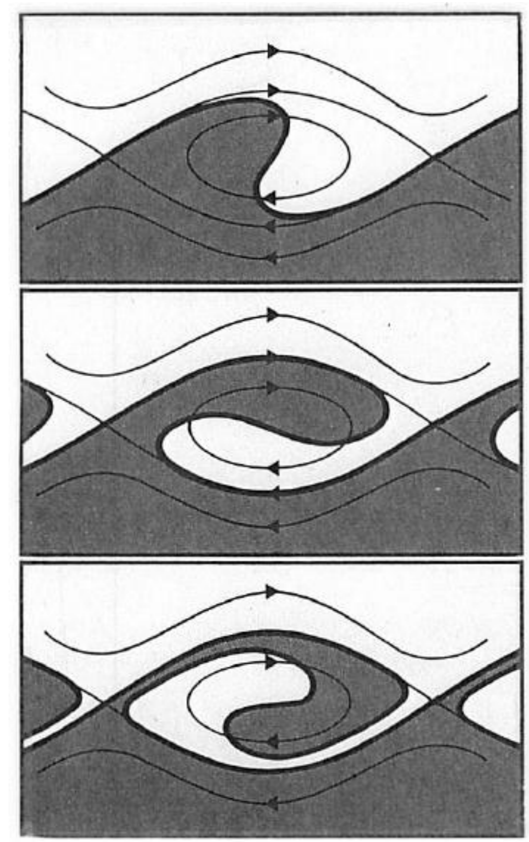


Postel and Hitchman 1999
Homeyer et al 2013

Rossby wave critical layer interactions (critical layer: U = c)



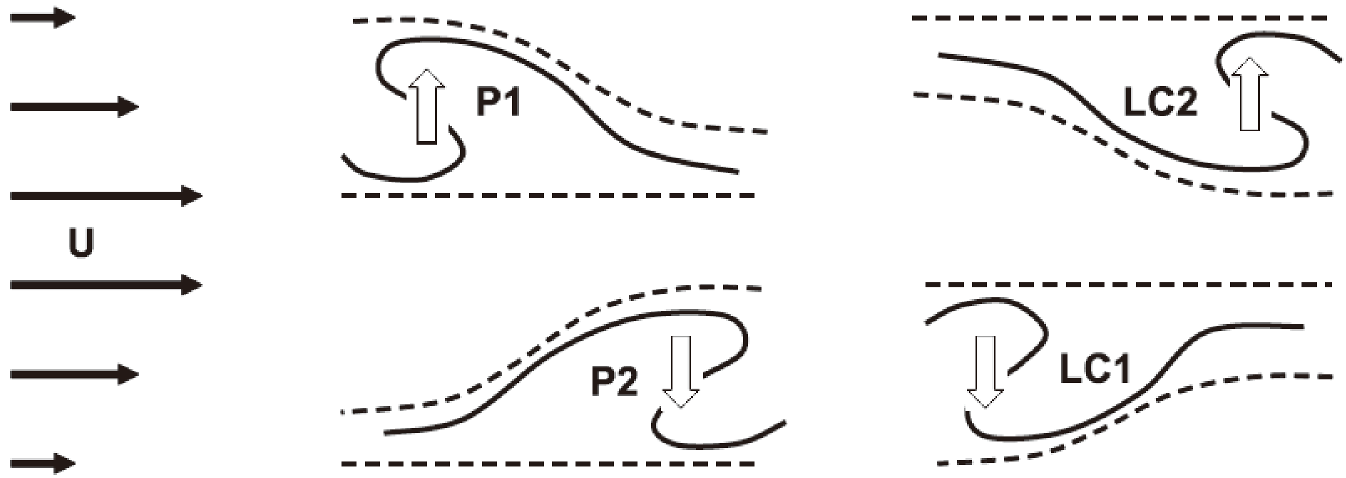
nonlinear overturning at critical layer
(irreversible transport and mixing)



Warn and Warn 1978

Two types of wave breaking, depending on shear of background winds

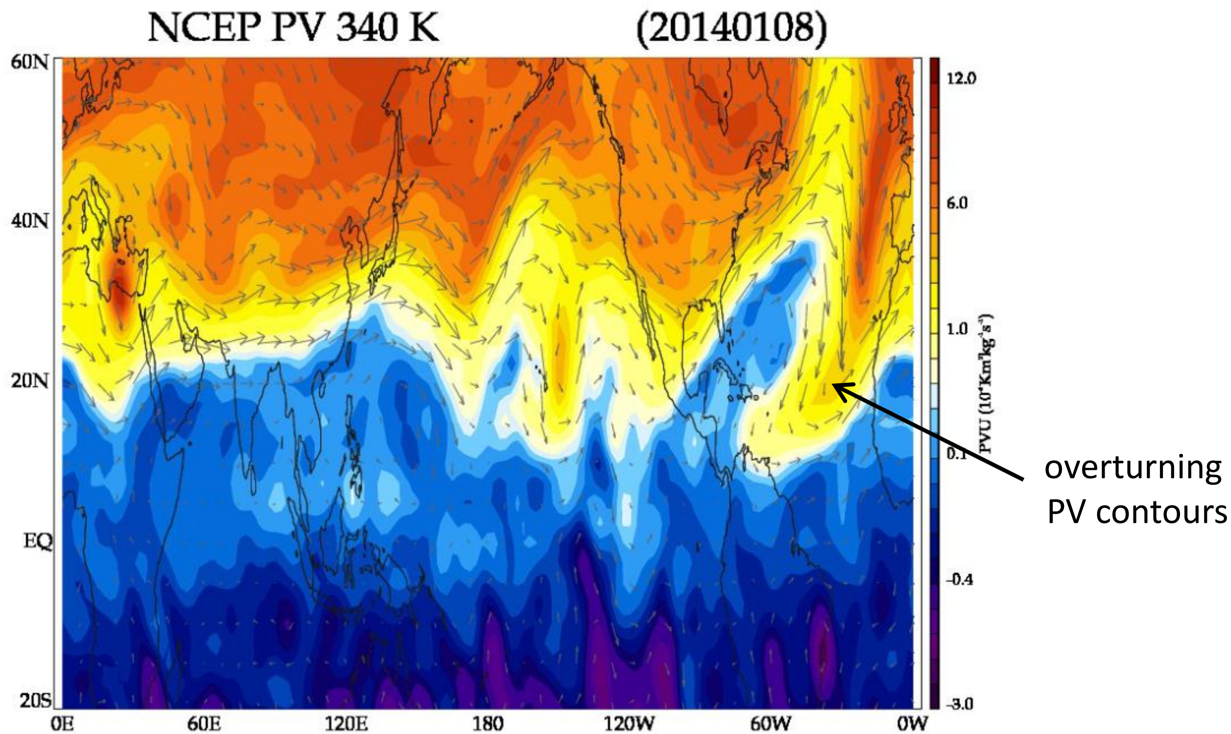
poleward breaking (cyclonic)



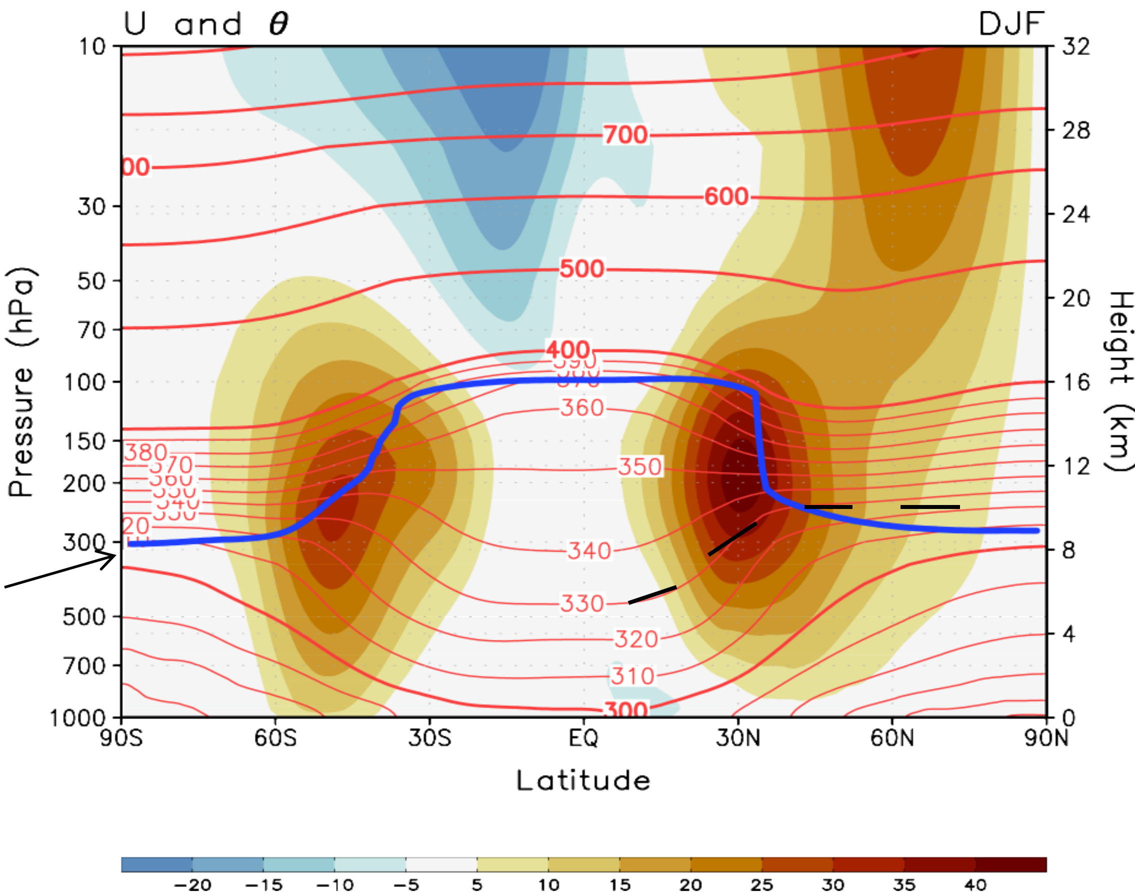
equatorward breaking (anticyclonic)

Gabriel and Peters 2008

Example of a large-scale breaking Rossby wave



fast, synoptic flow mainly along isentropes:



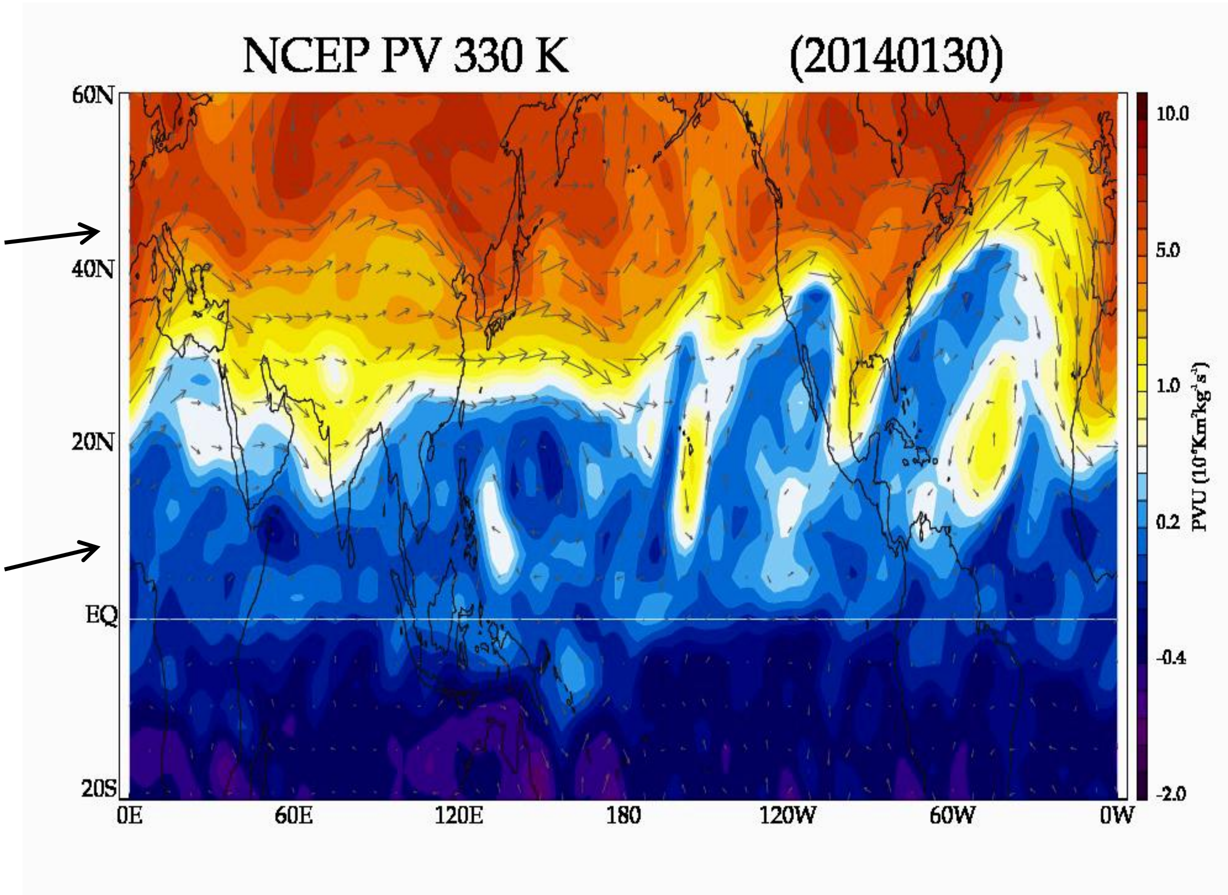
tropopause

330 K isentrope
spans tropical troposphere
to extratropical lower
stratosphere

Rossby waves during January-May at 330 K

extra-tropical lower stratosphere

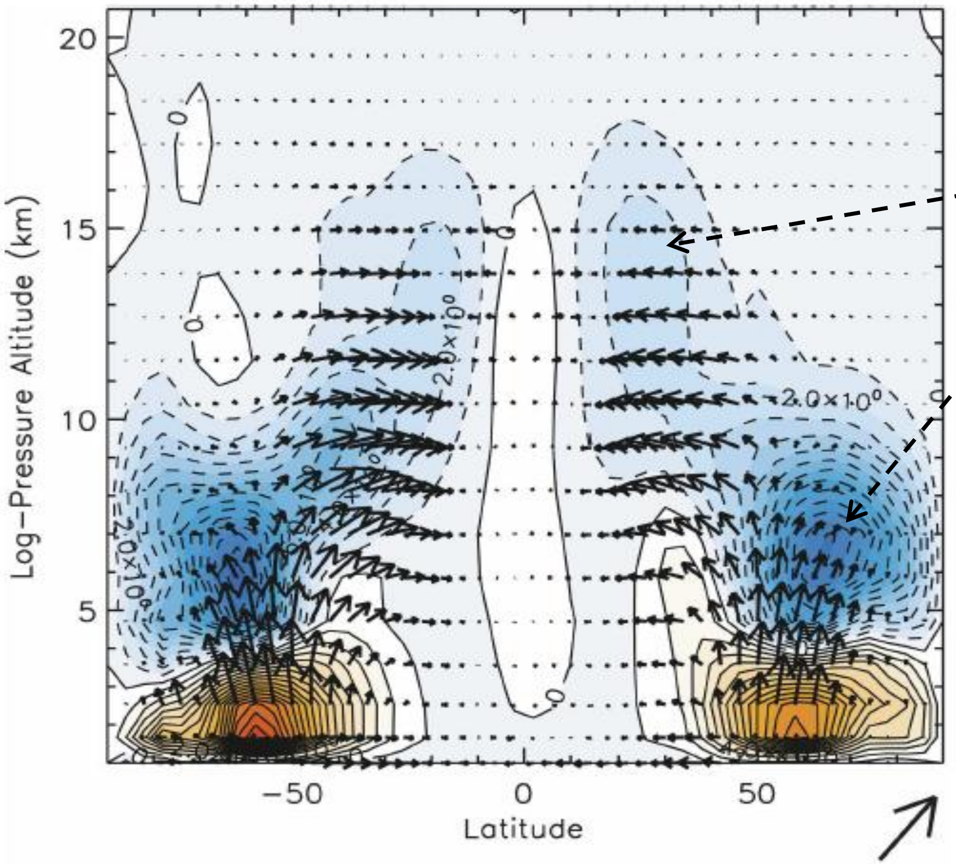
tropical troposphere



Key points:

- General refraction of Rossby waves towards low latitudes
- Latitudinal or vertical propagation for $U > 0$ (more generally $U > c$)
- Rossby wave breaking near critical lines ($U = c$)
- Poleward or equatorward breaking depending on background U shear
- Key mechanism for dissipation, mean flow forcing and transporting trace species

Climatological EP fluxes in the troposphere



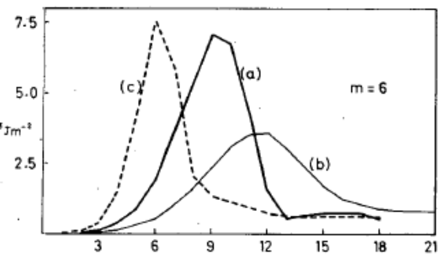
main regions of
wave dissipation

Garcia and Randel, JAS 2008

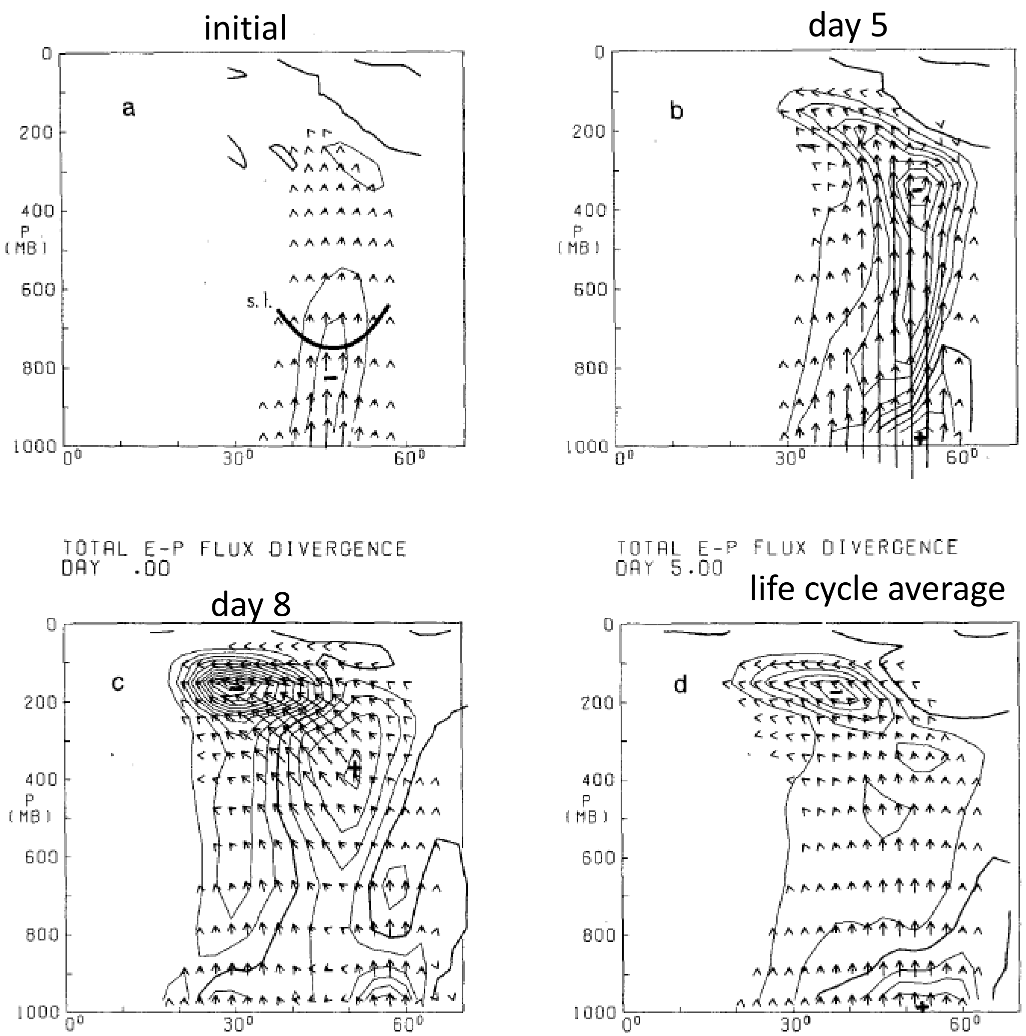
Extratropical EP flux patterns are related to baroclinic wave life cycles

idealized
zonal wave 6
baroclinic eddy
life cycle

energy



time (days)



baroclinic
growth

barotropic
decay

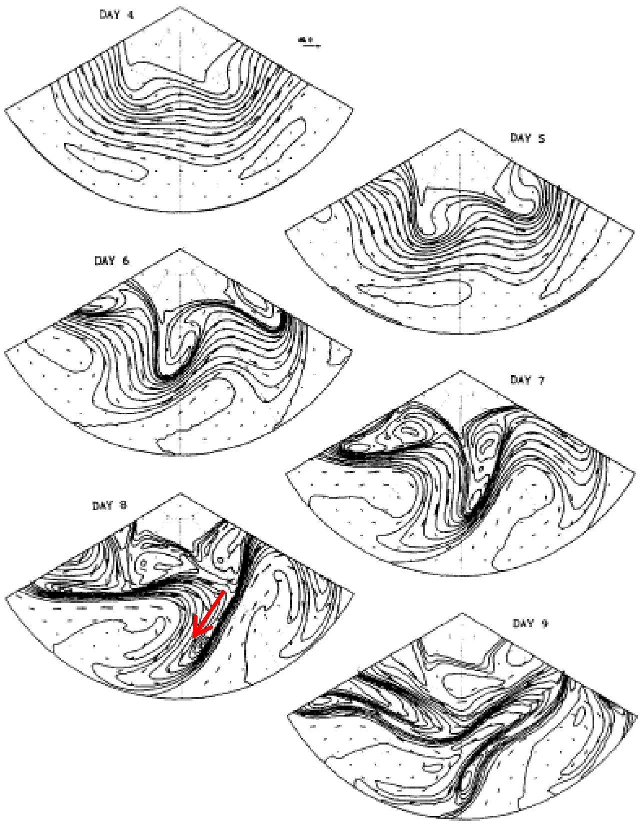
Simmons and Hoskins 1980
Edmon et al 1980

Q. J. R. Meteorol. Soc. (1993), **119**, pp. 17–55 551.511.32

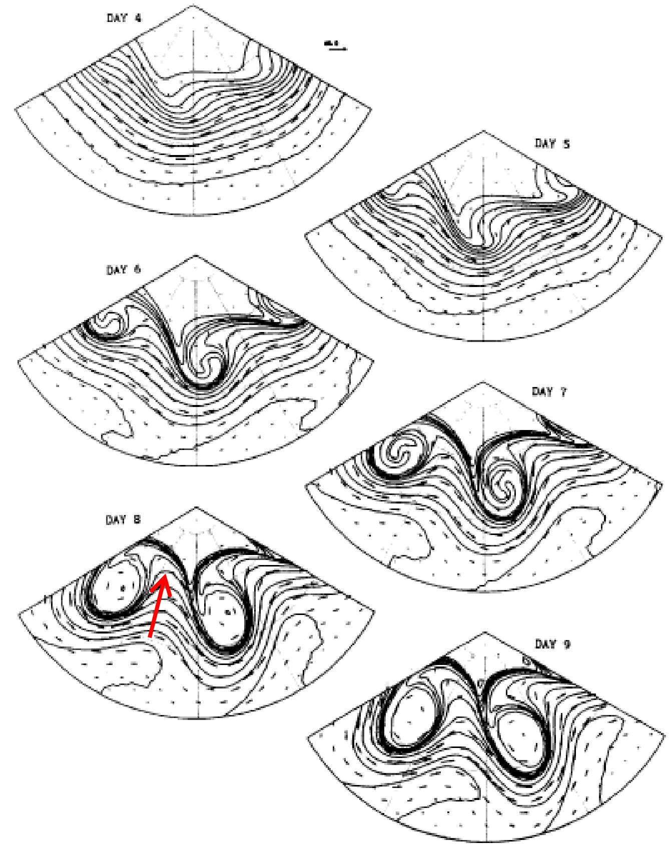
Two paradigms of baroclinic-wave life-cycle behaviour

By C. D. THORNCROFT[1*], B. J. HOSKINS[1] and M. E. McINTYRE[2]
[1] *Department of Meteorology, University of Reading*
[2] *Department of Applied Mathematics and Theoretical Physics, University of Cambridge*

LC1 equatorward breaking

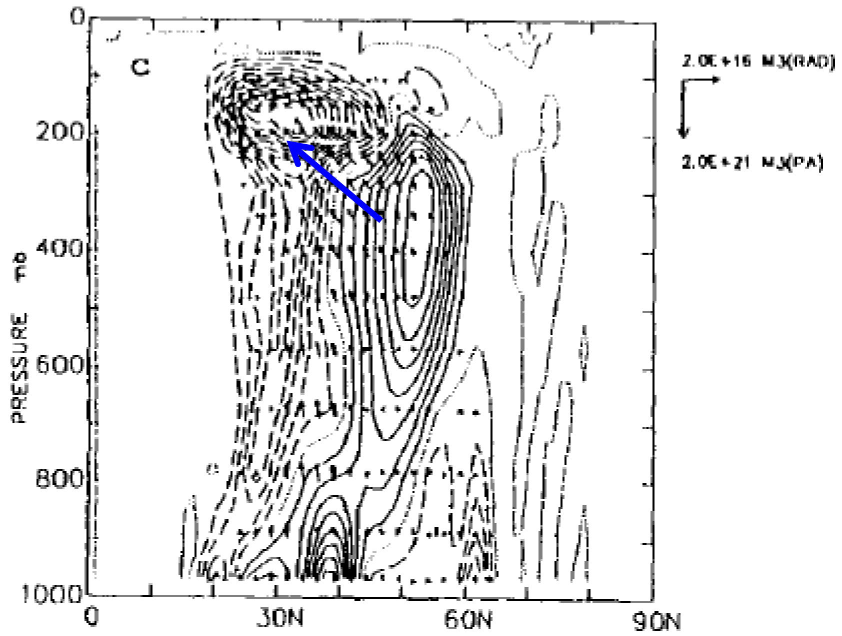


LC2 poleward breaking

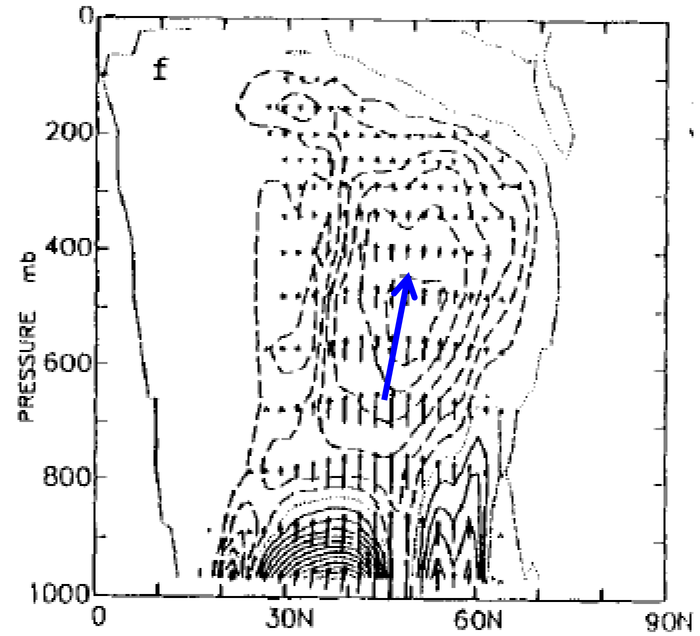


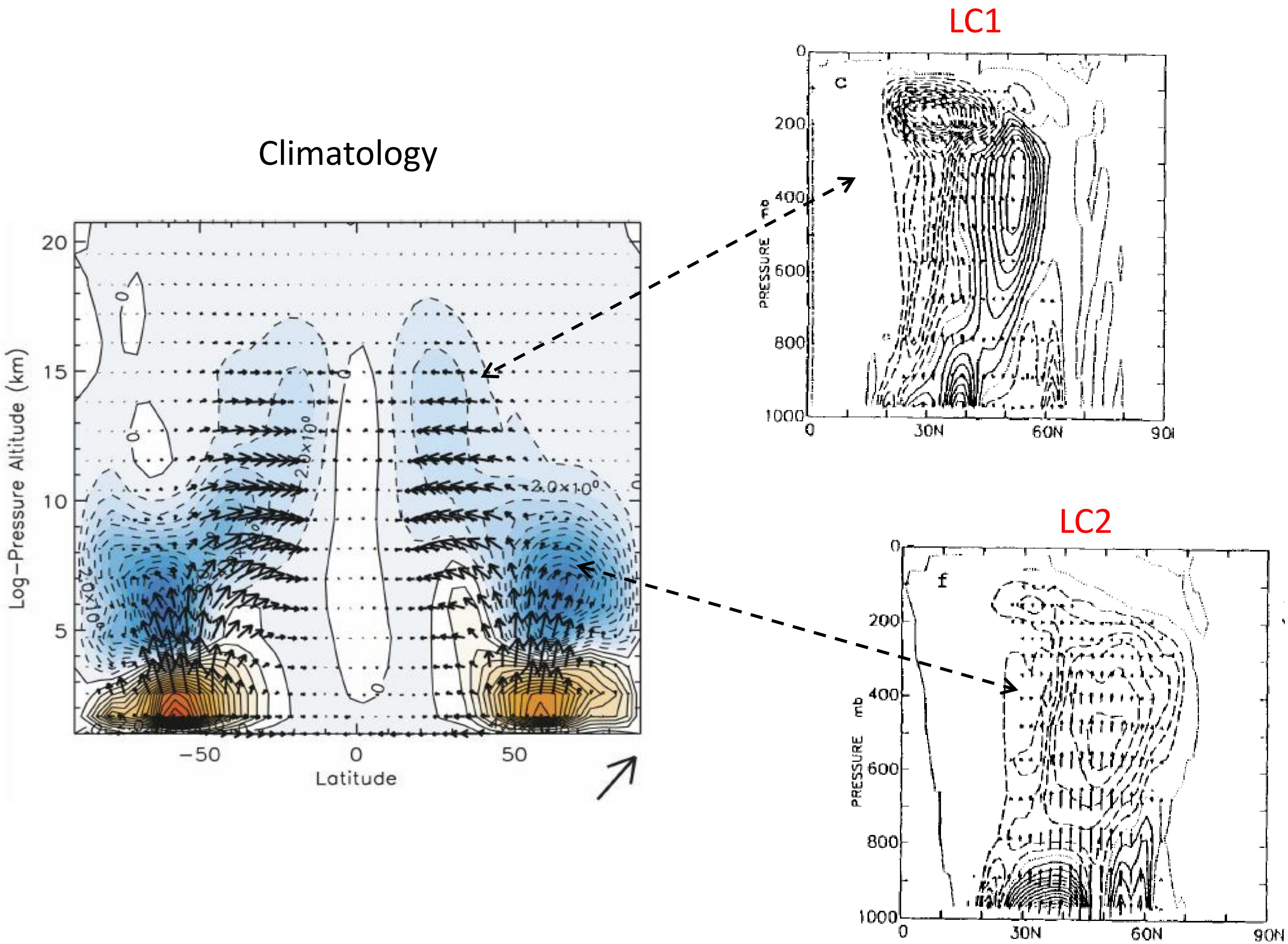
Idealized baroclinic wave life cycles

equatorward propagation (LC1)



poleward propagation (LC2)




Climatology
LC1
LC2
Log-Pressure Altitude (km)
Latitude
-50
0
50
2.0x10⁰
c
f
PRESSURE mb
0
200
400
600
800
1000
30N
60N
90N

Using phase speed spectra to diagnose critical layer interactions

wave flux
co-spectra as a function
of zonal wavenumber
and phase speed

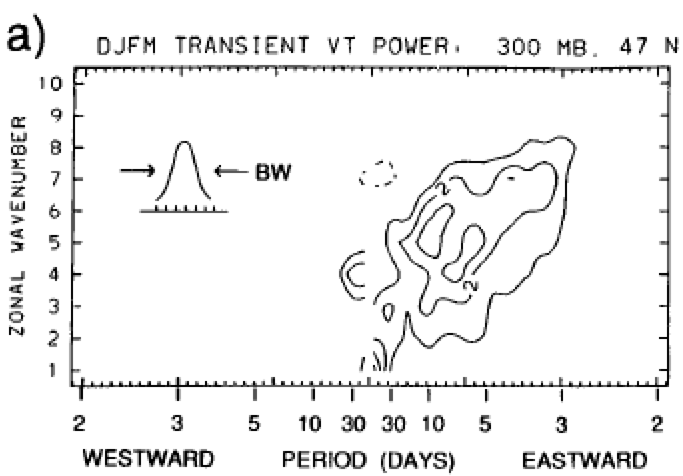
traditional
wavenumber vs.
frequency

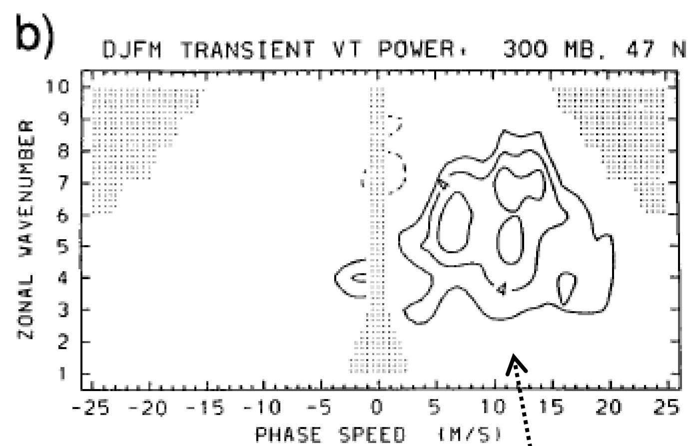
$$K_{n,c} = K_{k,\omega} \cdot \left(\frac{n}{a \cos \phi} \right).$$

Randel and Held 1991

wavenumber vs. frequency

wavenumber vs. phase speed

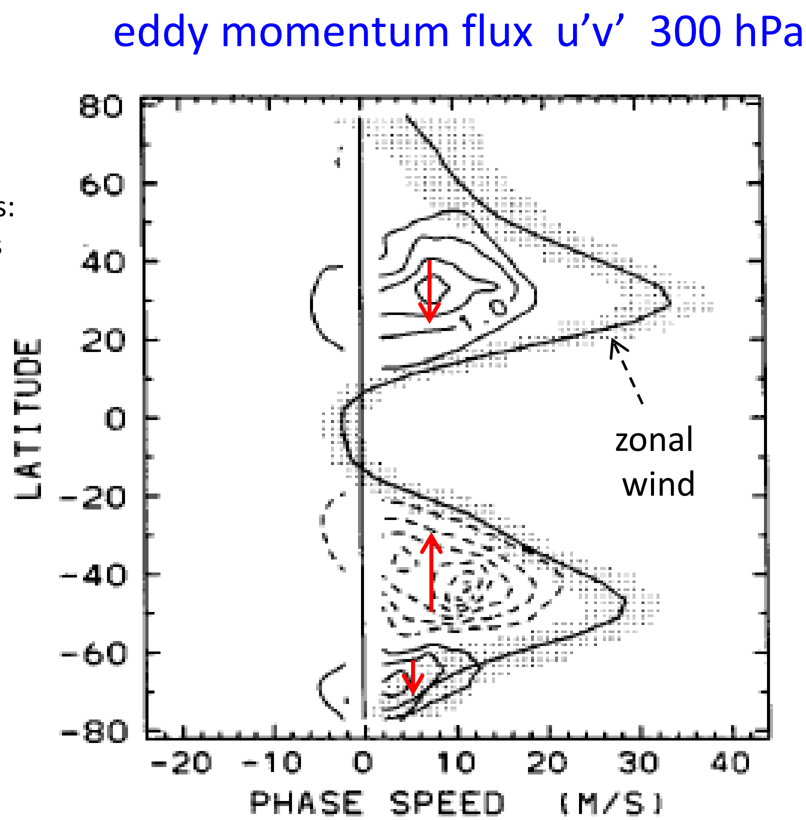




Rossby waves move eastward
at ~5-15 m/s

Integrate over wavenumber to derive eddy flux phase speed spectra

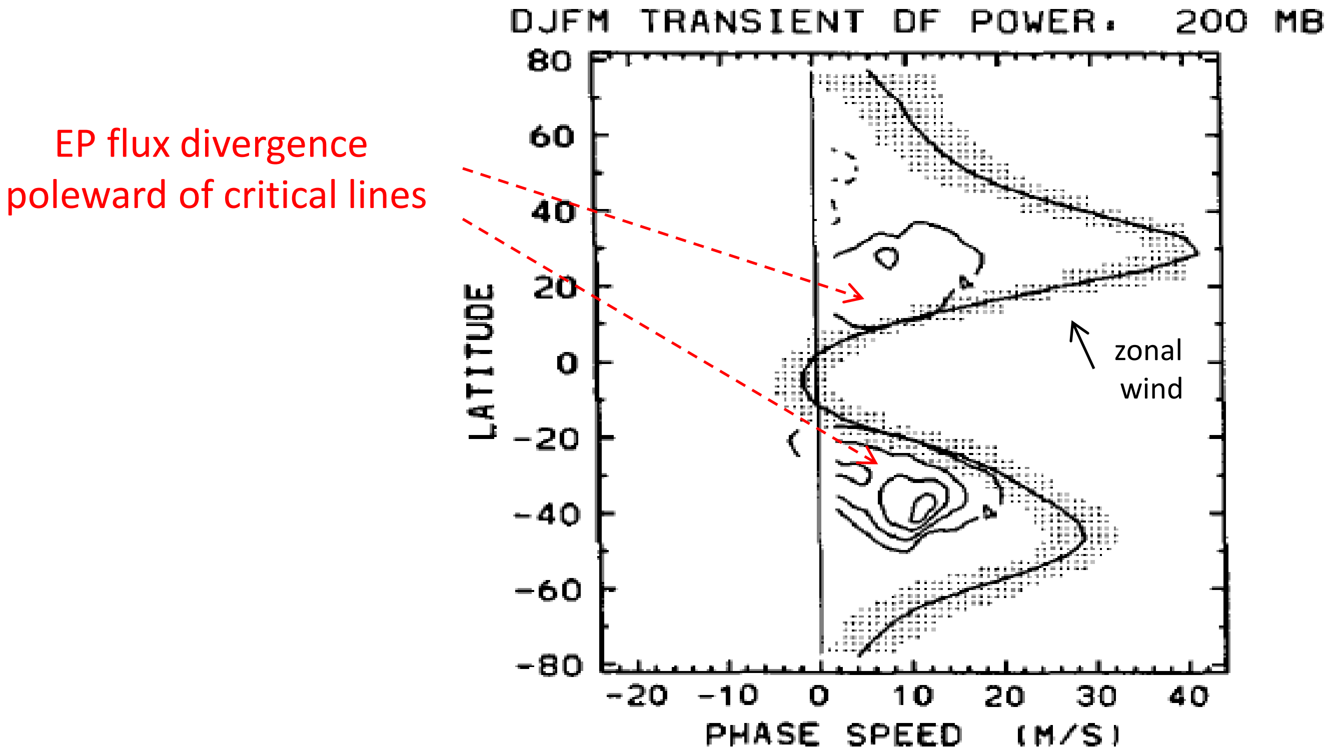
red arrows: EP fluxes



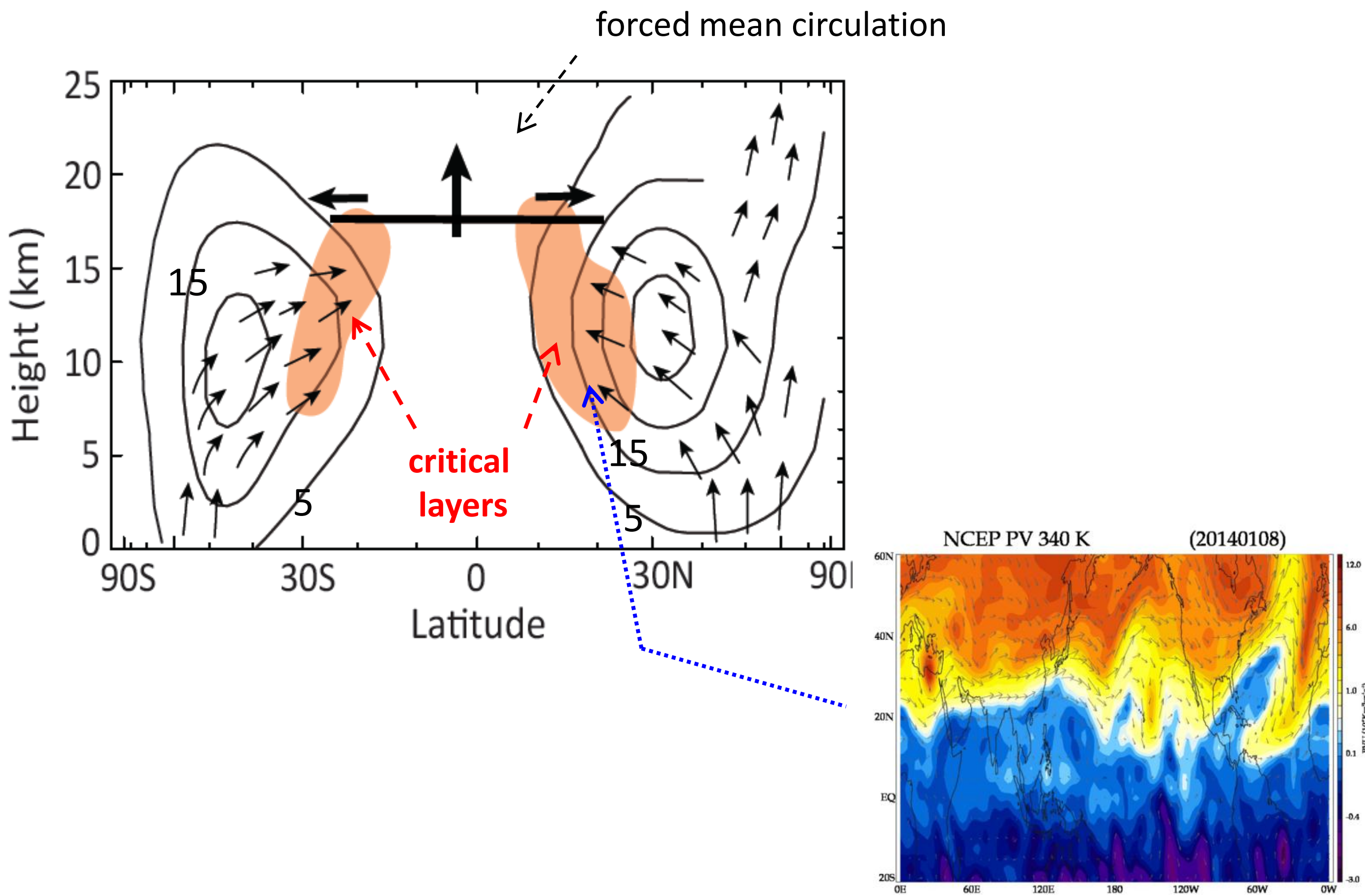
- EP fluxes: propagation to near critical lines (c = U)
- evidence for critical layer behavior

Randel and Held 1991

EP flux divergence phase speed spectra



Subtropical critical layers for Rossby waves with phase speeds ~ 5-15 m/s



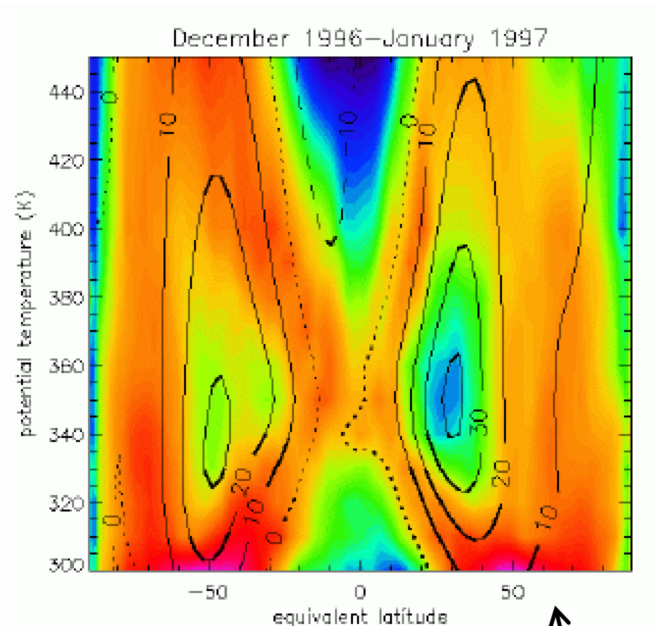
Effective diffusivity as a diagnostic of atmospheric transport
2. Troposphere and lower stratosphere

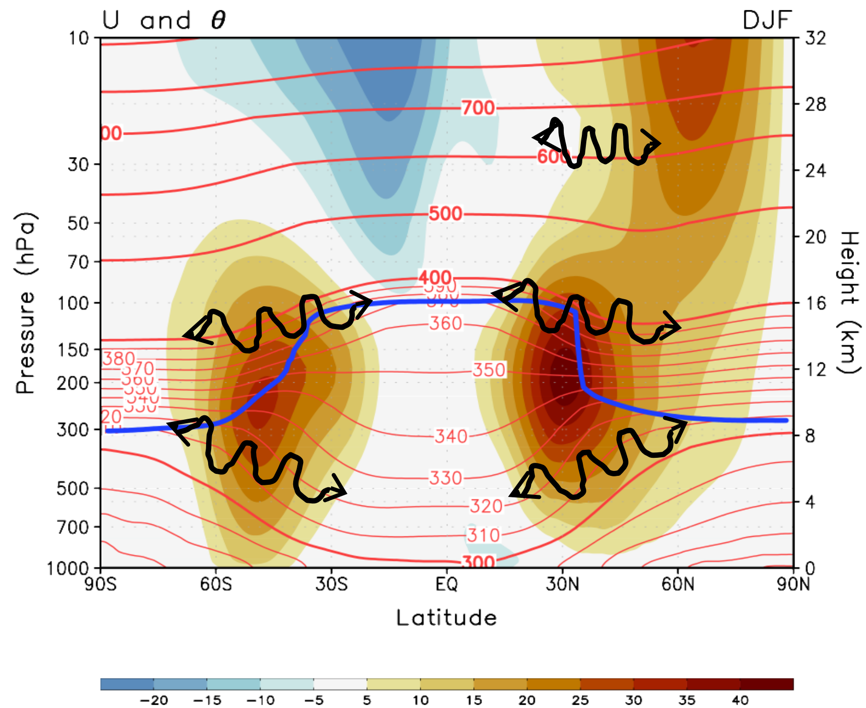
JGR 2000

Peter Haynes and Emily Shuckburgh

eddy transport above and below subtropical jets

Estimates of mixing based on stretching of PV contours in trajectory calculations





important points:

- mixing on flanks of jet (near critical lines for c ~10 m/s)
- small mixing across jet core (jet cores are mixing barriers)

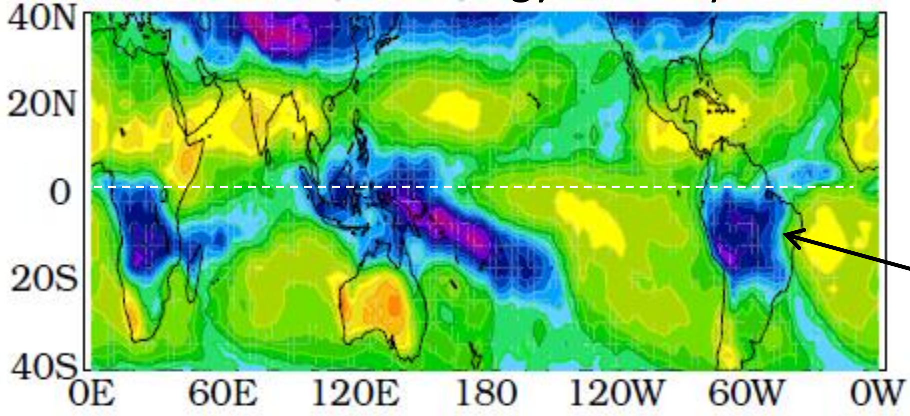
Key points:

- Baroclinic wave life cycles: baroclinic growth and barotropic decay
- Two idealized types of life cycles: equatorward and poleward wave breaking
  (LC1 and LC2)
- Consistent with tropospheric EP flux circulation statistics
- Phase speed spectra: clear evidence for critical layers in subtropics
  (important influence of extratropical waves on tropical circulations)

Large-scale tropical circulations are forced by latent heating from deep convection

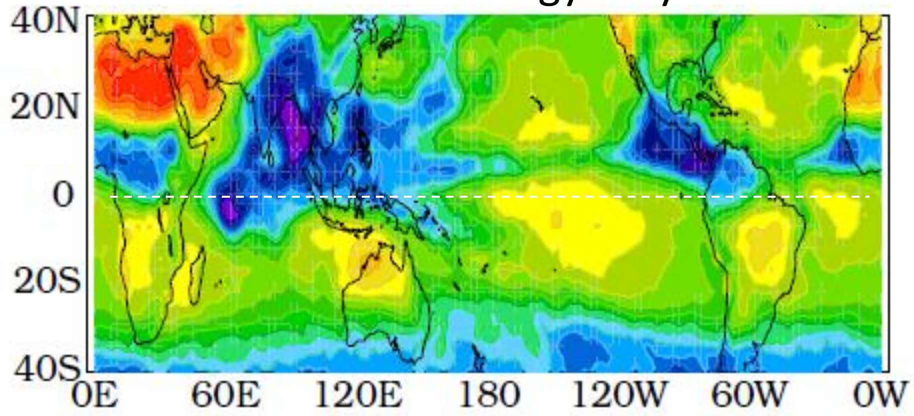
Outgoing Longwave Radiation (OLR) is a useful proxy for deep convection

OLR climatology January

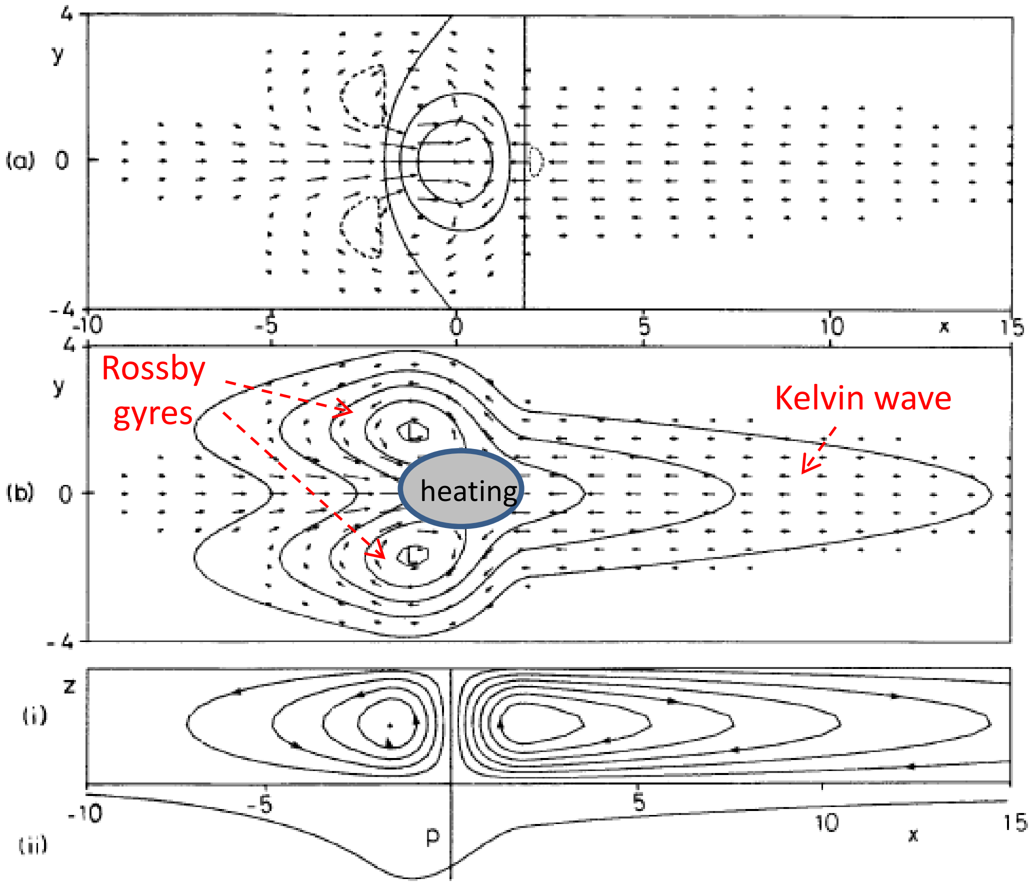


high clouds ~ deep convection

OLR climatology July



Dynamical response to low frequency convective forcing



low level convergence

low level circulation

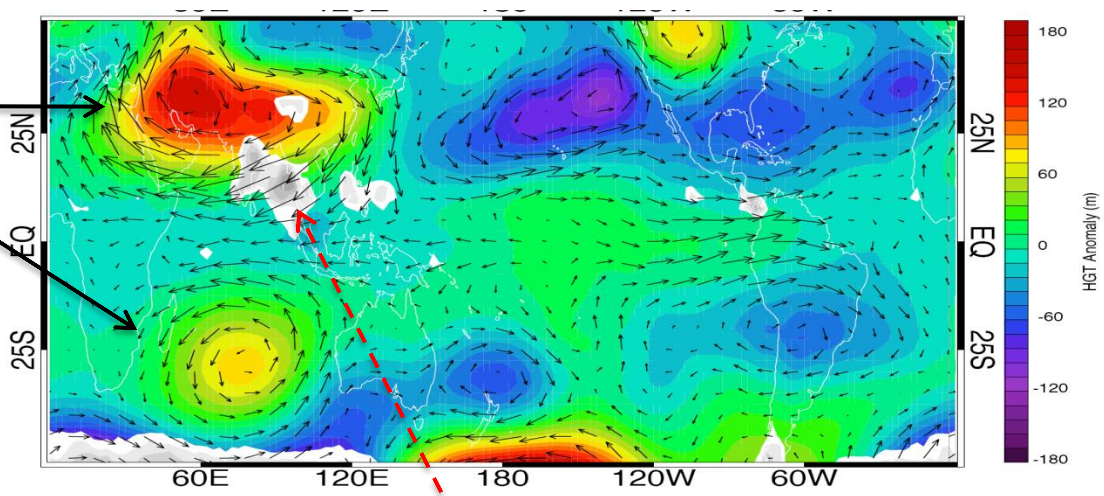
vertical structure

Gill, 1980

Tropical heating produces subtropical anticyclones in the UTLS

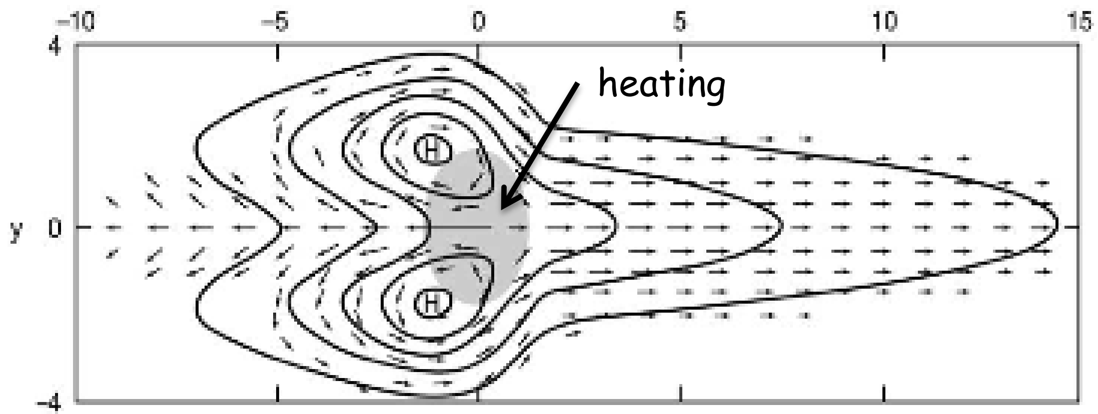
JJA geopotential height and winds 100 hPa

anticyclones



Convection (heating)

Park et al 2007



Matsuno-Gill Solution

Key points:

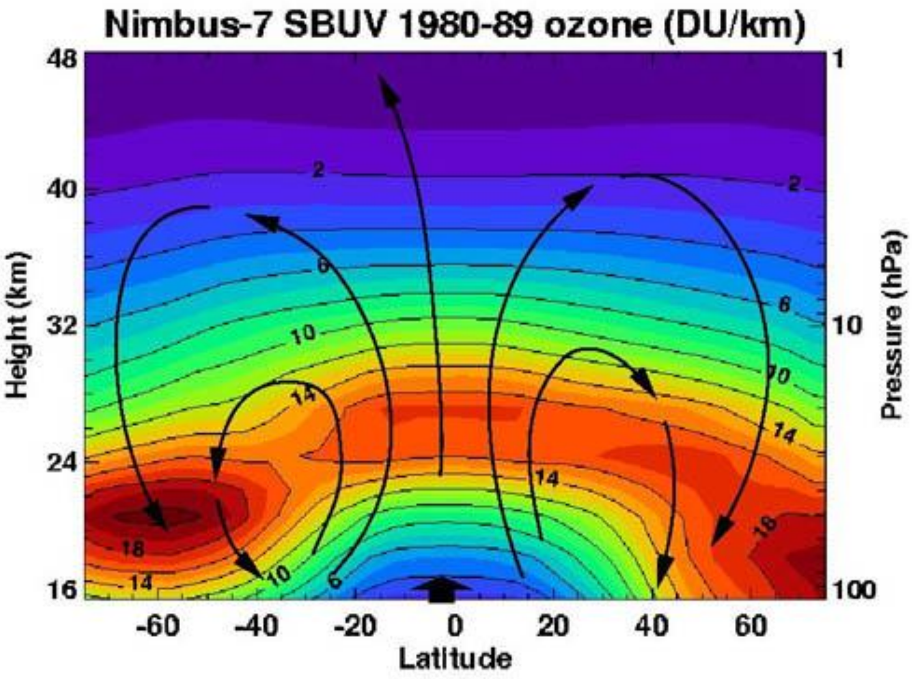
- Organized deep convection (latent heating) drives large-scale tropical circulations
- Seasonal movement between solstices (SH – NH subtropics)
- Matsuno-Gill dynamical response to local heating: subtropical Rossby waves and equatorial Kelvin waves
- Subtropical anticyclones in UTLS (especially Asian monsoon during NH summer)

Observed ozone and Brewer-Dobson circulation

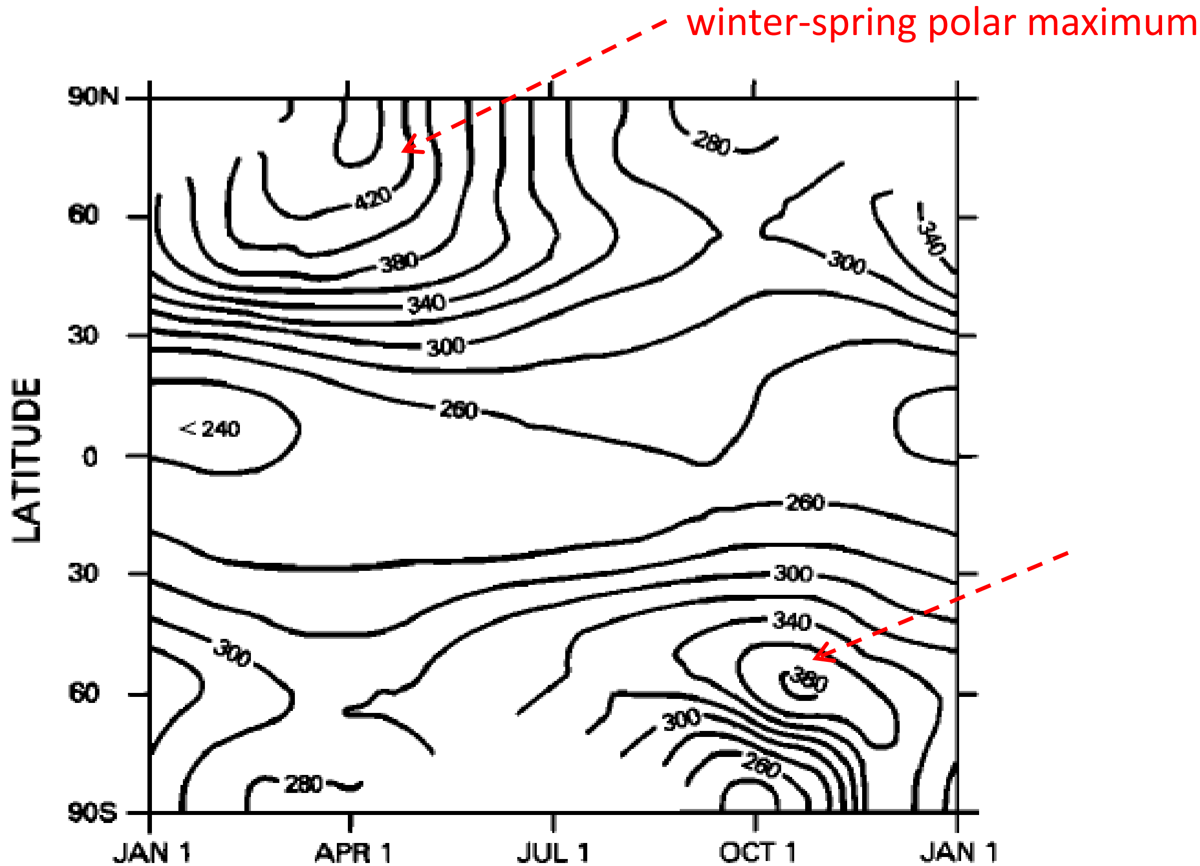
- ozone is made in the tropical stratosphere

- Short lifetime in upper stratosphere
- Long lifetime in lower stratosphere

- transport causes high latitude maximum
  during winter / spring



Seasonal cycle of column ozone reflects Brewer-Dobson circulation

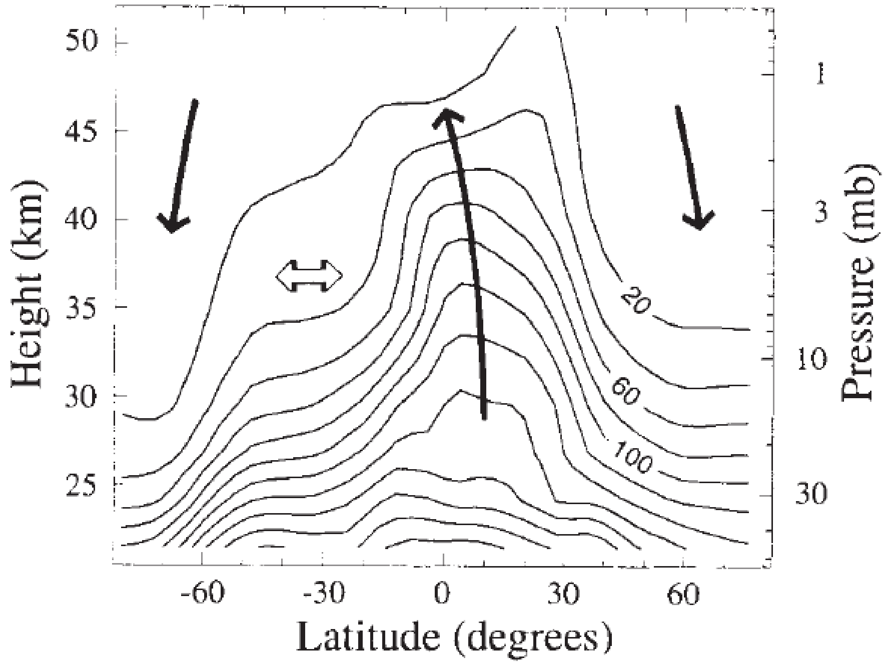


Bowman and Krueger, 1982

Stratospheric tracer transport: observations from satellites

- N_2O is a 'tropospheric source gas'
- destroyed by photolysis (radiation) in upper stratosphere
- Source of reactive nitrogen (NOx) in upper stratosphere; important for stratospheric ozone
- Behavior reflects Brewer-Dobson circulation and eddy mixing

N_2O mixing ratio



UARS observations from 1992

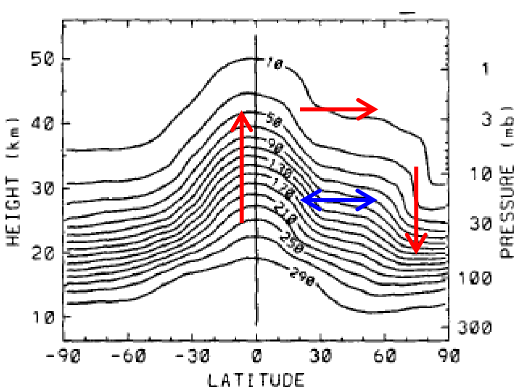
tracer zonal mean transport budget

mean advection

eddy transport

$$\frac{\partial \overline{\chi}}{\partial t} = -\overline{v}^{*} \frac{1}{a} \frac{\partial \overline{\chi}}{\partial \phi} - \overline{w}^{*} \frac{\partial \overline{\chi}}{\partial z} + \nabla \cdot \boldsymbol{M} + P - L$$

model N_2O (ppbv)

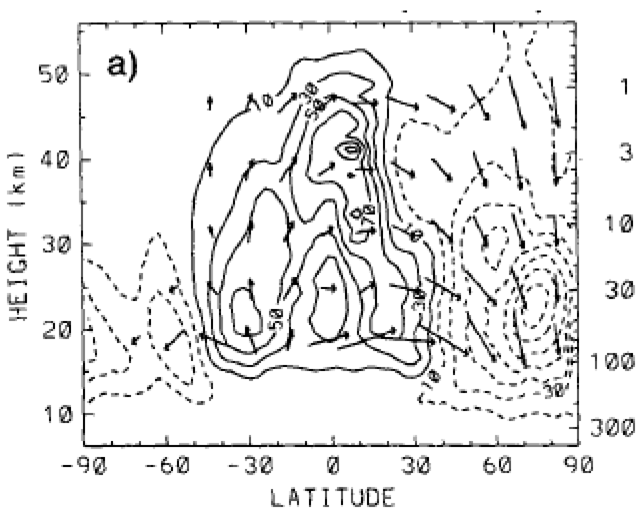


$$M_y = -e^{-z/H}\left(\overline{v'\chi'} - \frac{\overline{v'T'}}{S}\bar{\chi}_z\right)$$
$$M_z = -e^{-z/H}\left(\overline{w'\chi'} + \frac{\overline{v'T'}}{S}\bar{\chi}_y\right)$$

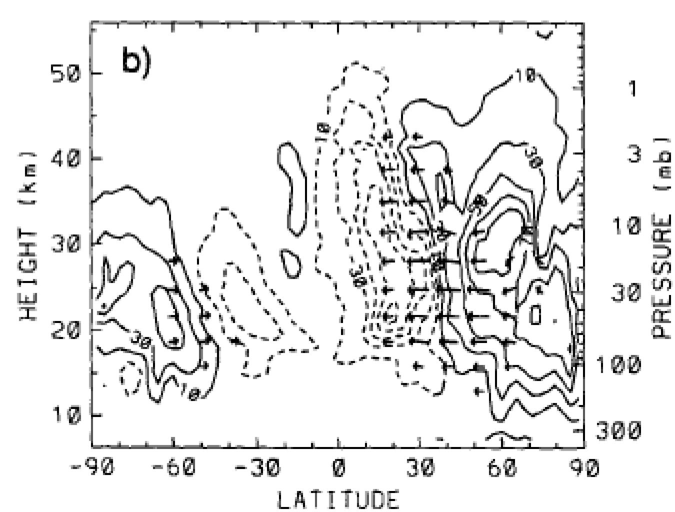
these terms generally balance each other

contours: N_2O tendency (ppbv/100 days)

mean advection



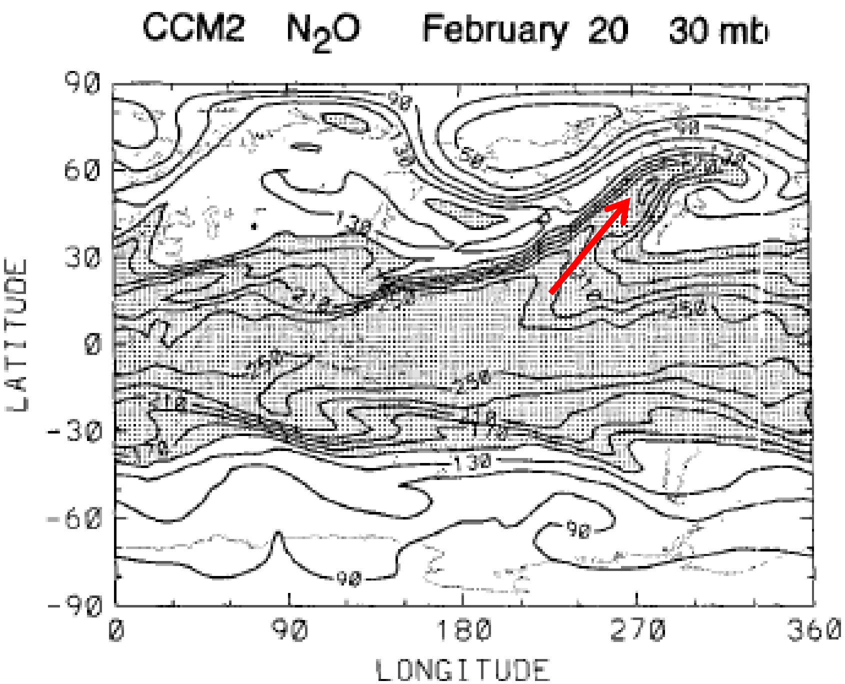
eddy transport



Randel et al 1994

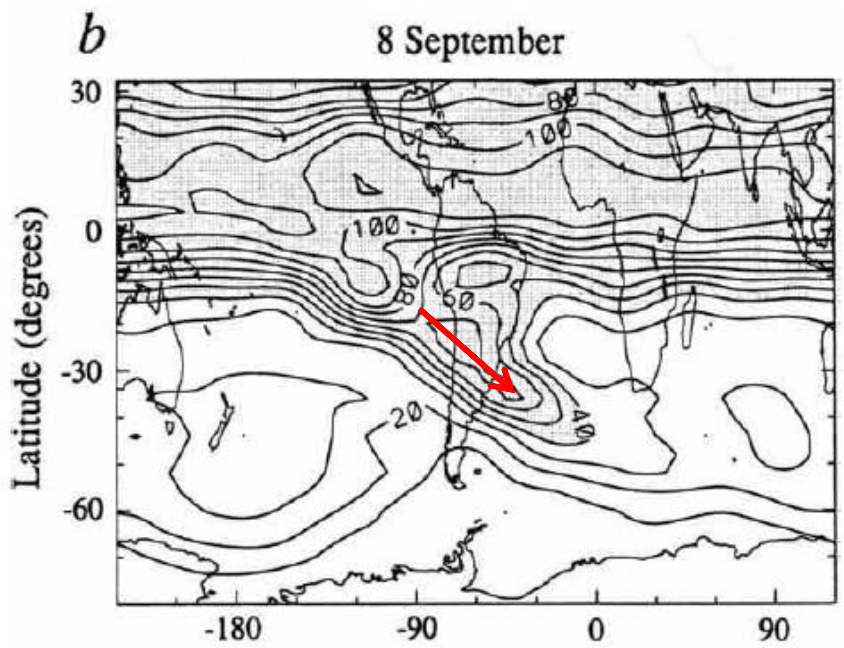
Examples of stratospheric wave mixing

model



N_2O near 35 km from CLAES instrument on UARS

observations



Randel et al 1993

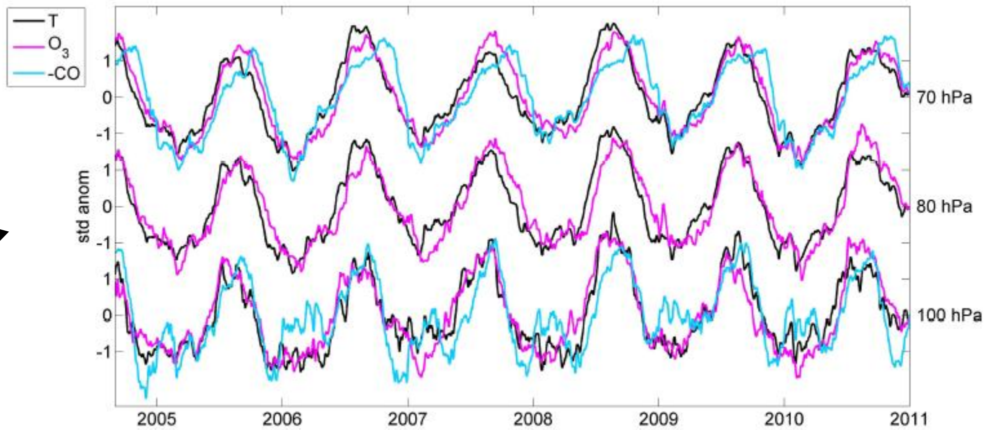
Tracer transport equation similar to thermodynamic equation:

tracer
$$\frac{\partial \overline{\chi}}{\partial t} = -\overline{v}^* \frac{1}{a} \frac{\partial \overline{\chi}}{\partial \phi} - \overline{w}^* \frac{\partial \overline{\chi}}{\partial z} + \nabla \cdot \boldsymbol{M} + P - L$$

temperature
$$\frac{\partial \overline{T}}{\partial t} + \overline{v}^* \frac{1}{a} \frac{\partial \overline{T}}{\partial \phi} + \overline{w}^* S = \overline{Q},$$

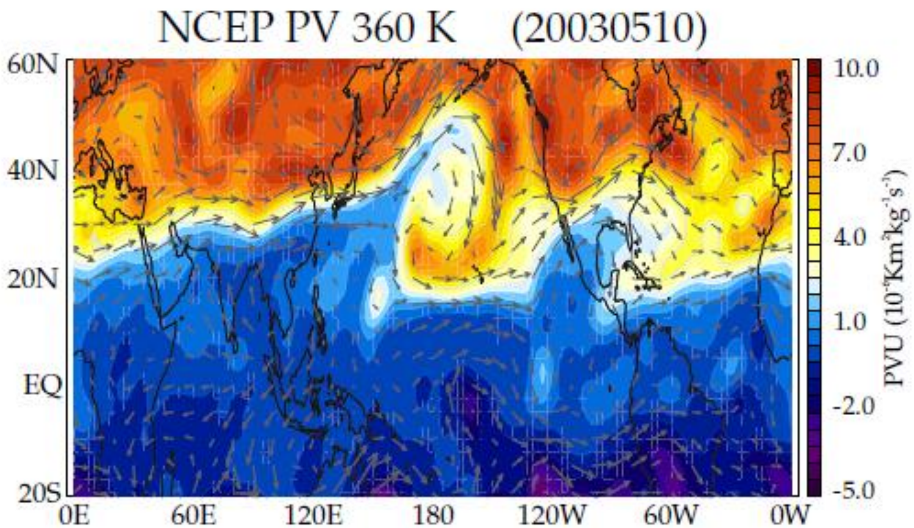
This is why temperature and tracers are sometimes highly correlated:

for example,
T, O_3 and CO
in tropical
stratosphere
(Abalos et al 2012)

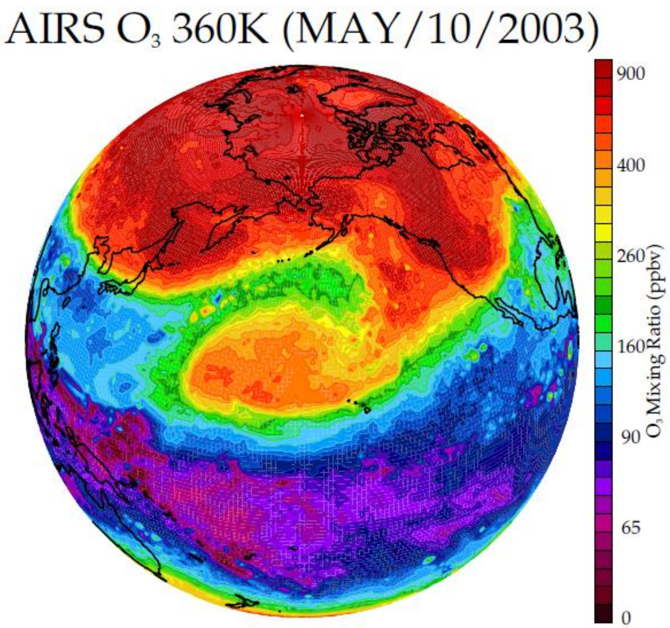


Mixing across tropopause linked to Rossby wave breaking

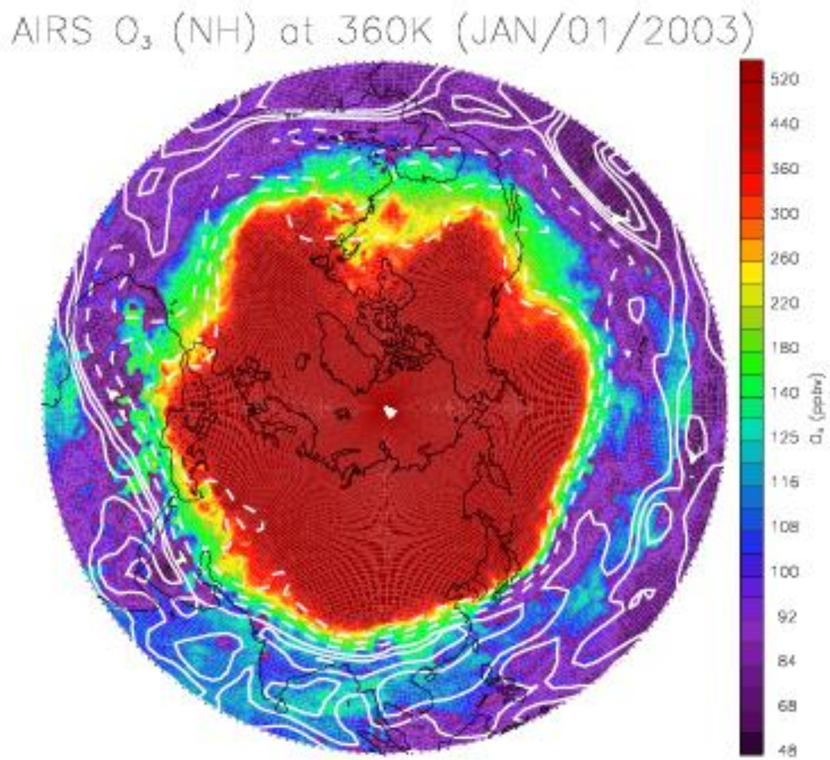
potential vorticity



ozone derived from AIRS



Rossby wave variability reflected in ozone near tropopause



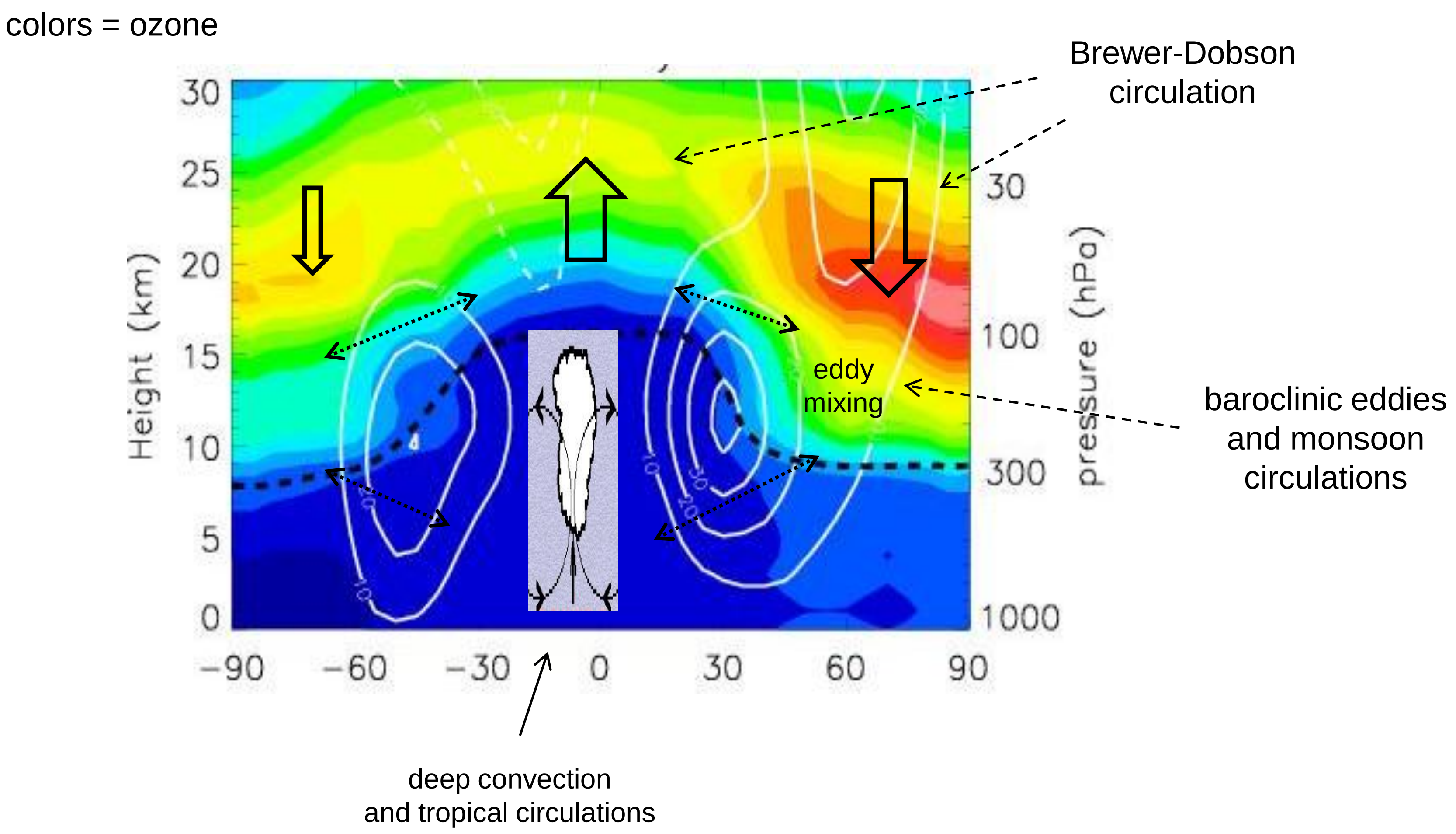
Key points:

- Stratospheric transport: Brewer-Dobson circulation and wave mixing

- Stratrospheric ozone: produced in tropical stratosphere, transported to high latitudes (reflects seasonal Brewer-Dobson circulation)

- Tracer budgets: mean advection and eddy transports (tied to Rossby waves and critical layers)

UTLS circulation and transport




Thank You