Mechanisms for stratospheric influences on tropospheric climate

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Outline

- The stratosphere
  - → Nature of variability
  - → Long-term memory
  - → Some open issues
- The "upper boundary condition" for the troposphere
- The troposphere
  - → Tropics and extratropics
  - → EPFD and the STJ
  - → Lessons from LC1/LC2
  - → Stratospheric impact on tropospheric NAM/SAM?

The stratosphere

- Balance between radiative relaxation (with seasonal cycle) and wave-driven circulation: $\dfrac{d\theta}{dz}\,\overline{w}^* \approx -r(\theta - \theta_{\rm rad})$

- Circulation largely wave-driven, forced by convergence of angular pseudomomentum (equivalently EPFD)

  → Also forced by seasonal cycle (Garcia 1987 *JAS*)

  → Local perturbations by waves (even if non-dissipating)

- System is energetically open (because of cooling to space)

  → Diabatic heating or cooling cannot be imposed (Fels 1985 *Adv. Geophys.*)

  → Exception is middle atmosphere Hadley circulation (Dunkerton 1989 *JAS*)

- Angular momentum budget is tightly constrained (no external sources or sinks)

  → Circulation is largely mechanically driven, thermally damped (refrigerator)

  → Illustrated by response to imposed force or radiative perturbation (Haynes *et al.* 1991 *JAS*)

- Short-term variability (apart from seasonal cycle) in zonal-wind/temperature structure is driven by — and can *only* be driven by — variability in EPFD

  → Radiative-chemical behaviour is *dissipative*; chaos arises only from dynamics

  → Same argument applies to interannual variability (except for volcanoes and solar variability)

  → However there is coupling between dynamics and chemistry (which may enhance or skew variability)

- Variability in EPFD can arise from variability in tropospheric wave sources, or internally within the stratosphere

- Wintertime stratosphere has "strong vortex" and "weak vortex" states (Holton & Mass 1976 *JAS*)

  → Correspond to extreme states of the NAM and SAM

  → Involves stratospheric role in stratospheric variability (Christiansen 2000 *JAS*; Scott & Haynes 2000 *JAS*)

  → Must be linked to EPFD (by angular momentum)

- Possible nonlinear feedback: a weaker vortex allows more EPFD which further weakens it (and vice-versa)

  → Seen in sudden warmings (Dunkerton *et al.* 1981 *JAS*)

  → How robust is this relationship?

- Stratosphere has considerable long-term memory because of the "low-latitude flywheel" (Scott & Haynes 1998 *QJRMS*)

  → Equatorial zonal wind anomalies are only weakly damped by radiation

  → Leads to interannual variability in stratosphere-only models (Scott & Haynes 1998 *QJRMS*)

  → Reflected(?) in inter-hemispheric correlations in variability of midlatitude total ozone (Fioletov & S. 2003 *GRL*)

- Summertime stratosphere is relatively quiescent

  → Reflected in persistence of midlatitude total ozone anomalies (Fioletov & S. 2003 *GRL*)

  → And in persistence of zonal wind anomalies

  → How well is this persistence represented in GCMs?

  → Role of internal instability (Norton 2002 *GRL*)?

Some (other) loose ends

- Focus is usually on planetary-wave drag (PWD) as the driver of the stratospheric circulation. To what extent is this OK?
  - → Evidence for role of synoptic-scale wave drag (SSWD) (Norton 2001 Bad Tölz workshop)
  - → Importance of gravity-wave drag (GWD) (Garcia & Boville 1994 *JAS*; Beagley *et al.* 1997 *Atmos.–Ocean*)
- How important is GWD in the response to a PWD anomaly?
  - → How well is this represented in GCMs?
  - → Drag schemes which are not momentum-conserving may give spurious responses (S. *et al.* 1996 *JGR*)
- What is the PWD (and maybe GWD) response to a radiative perturbation?
  - → Kiehl *et al.* (1988 *Nature*) suggest a colder vortex leads to reduced PWD (positive feedback), but is this robust?
  - → A major issue for GHG effects on polar ozone (albeit only part of the problem)
  - → Drag schemes which are not momentum-conserving may give spurious responses (S. *et al.* 1996 *JGR*)

- Need to distinguish within-stratosphere PWD effects from changes in PW forcing from the troposphere

- Some "small" effects of the stratosphere on tropospheric climate

  → Radiative transfer

  → Actinic fluxes, which affect tropospheric ozone

  → Dynamically induced upwelling and downwelling

- But how small are they, in practice?

  → The relative mass of the troposphere argues against these direct non-local effects being very significant

  → But in a nonlinear system, such an argument is dangerous (e.g. via tropopause height: Thuburn & Craig 2000 *JAS*)

- Stratosphere provides an upper boundary condition for the troposphere

  → Affects stationary waves

  → But the nature of this upper boundary condition has not been much investigated

  → Boville & Cheng (1988 *JAS*), but what since?

  → Seen this way, the "mass" argument is inapplicable

The troposphere

- Tropics (< 11 km): Convectively adjusted; Hadley circulation
  - → Balance between convective heating and clear-sky cooling
  - → In tropical stratosphere, balance is instead between mean upwelling and clear-sky heating
- Tropical tropopause layer (TTL): transition region between troposphere and stratosphere (Thuburn & Craig 1998 *JAS*)
  - → How does the stratosphere affect the TTL?
  - → How does the TTL affect the lower tropical troposphere?
- Arguments on the control of the convective outflow (and thus cloud-top temperatures) differ:
  - → Folkins (2002 *JAS*) argues outflow level set by distribution of Θ_e in boundary layer and stratification in TTL
  - → Hartmann & Larson (2002 *GRL*) argue that the outflow temperature is set by the Clausius-Clapeyron relation
- Extratropics: Baroclinicly adjusted
  - → As in stratosphere, represents balance between EPFD (including surface processes) and radiation

→ To what extent is radiation relaxational (since clouds are involved)?

- Connection between tropics and extratropics via STJ not much studied, nor well understood

- In stratosphere, PWD is *forced* (paradigm is Rossby-wave critical layer)

  → In extratropical troposphere, EPFD is associated with SSWD and involves *instability*

  → Suggests much tighter interaction with the mean flow

  → The baroclinic adjustment problem remains poorly understood

- Strength and location of subtropical jet (STJ) only weakly constrained (as with stratospheric polar night jet)

  → Variability of STJ (and AMs) associated with EPFD (Limpasuvan & Hartmann 2000 *JC*)

- Variability in EPFD mainly leads variability in AMs (Lorenz & Hartmann 2001 *JAS*, 2003 *JC*)

  → Basically this has to be the case

  → Possible feedback may not be well diagnosed this way

- Lorenz & Hartmann argue that a positive AM anomaly
  $\Rightarrow$ stronger baroclinicity
  $\Rightarrow$ stronger synoptic-scale momentum flux into jet

  → They also argue that low-frequency planetary-wave drag acts to weaken the jet (cf. S. 1987 *JAS*)

- Recall LC1/LC2 paradigm (Thorncroft *et al.* 1993 *QJRMS*)

  → Addition of cyclonic shear causes the SSWD to move from the equatorward side of the STJ (LC1) to the poleward side (LC2)

  → Reflected in decrease in poleward momentum flux

  → Will have impact on surface pressure and surface AMs (Haynes & S. 1989 *QJRMS*)

- There is evidence for the LC1/LC2 dichotomy in observations (Randel & Held 1991 *JAS*)

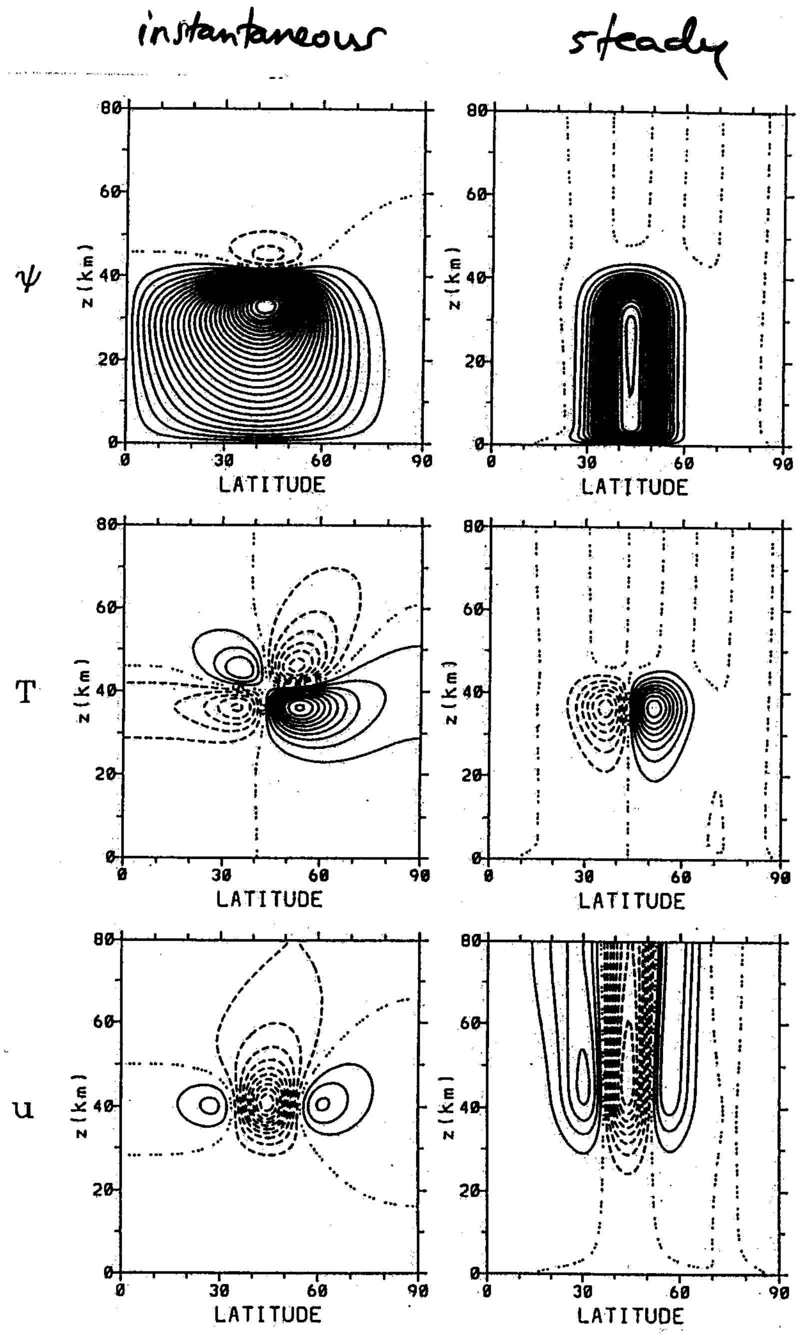
- Cyclonic shear corresponds to a stronger STJ and a negative AM anomaly (Baldwin, personal communication)

- One may thus hypothesize that a positive AM anomaly in the stratosphere will weaken the STJ and bias the troposphere towards more LC1-like behaviour, with a positive momentum flux anomaly

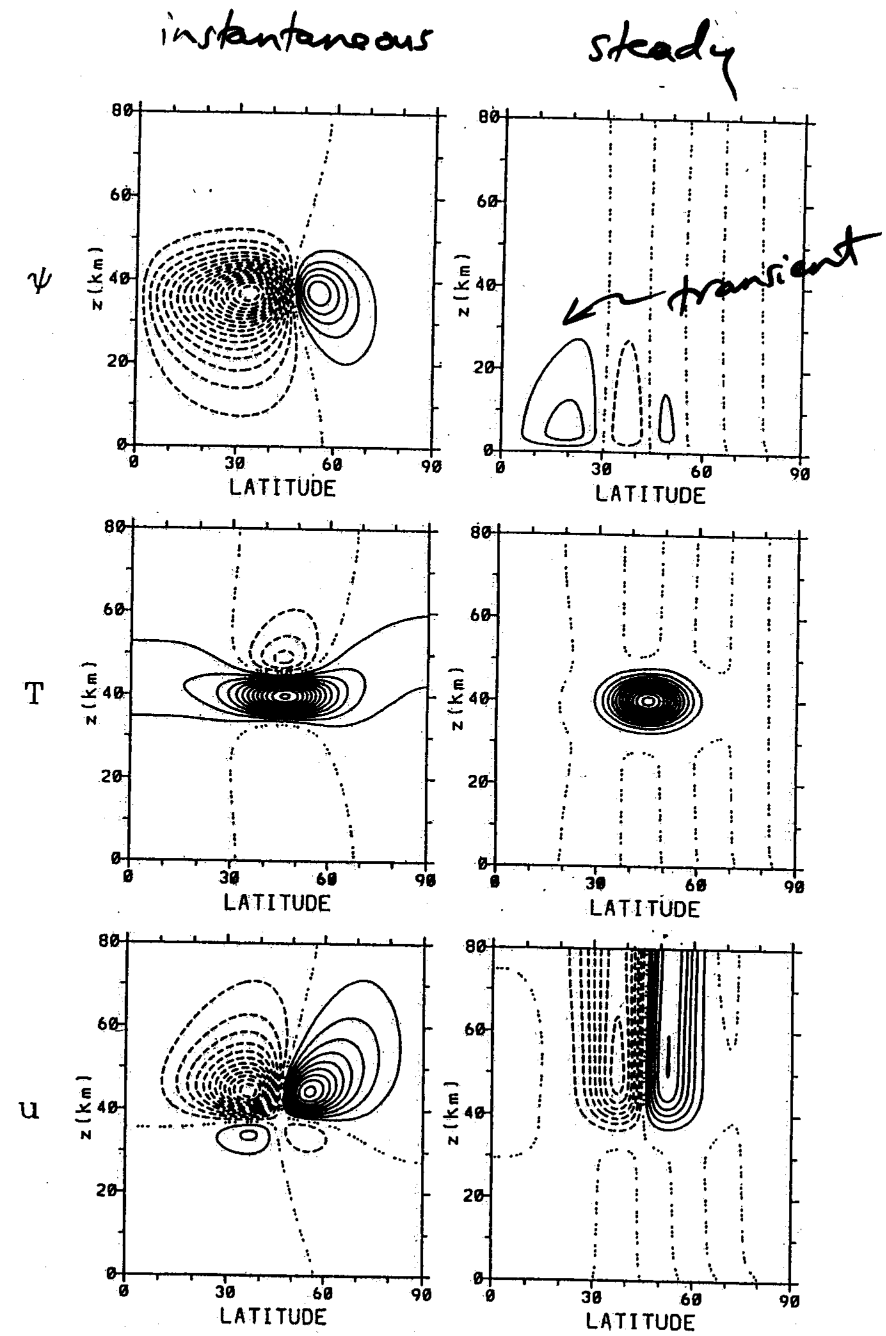
  → Seems to be true (Baldwin, personal communication)

Shepherd, Semeniuk & Koshyk (1996 JGR)

Response to imposed force



Response to imposed heating



Distribution of wintertime temperatures (86°N, 2.6 hPa) in millenial integrations with a mechanistic strat-trop PE model

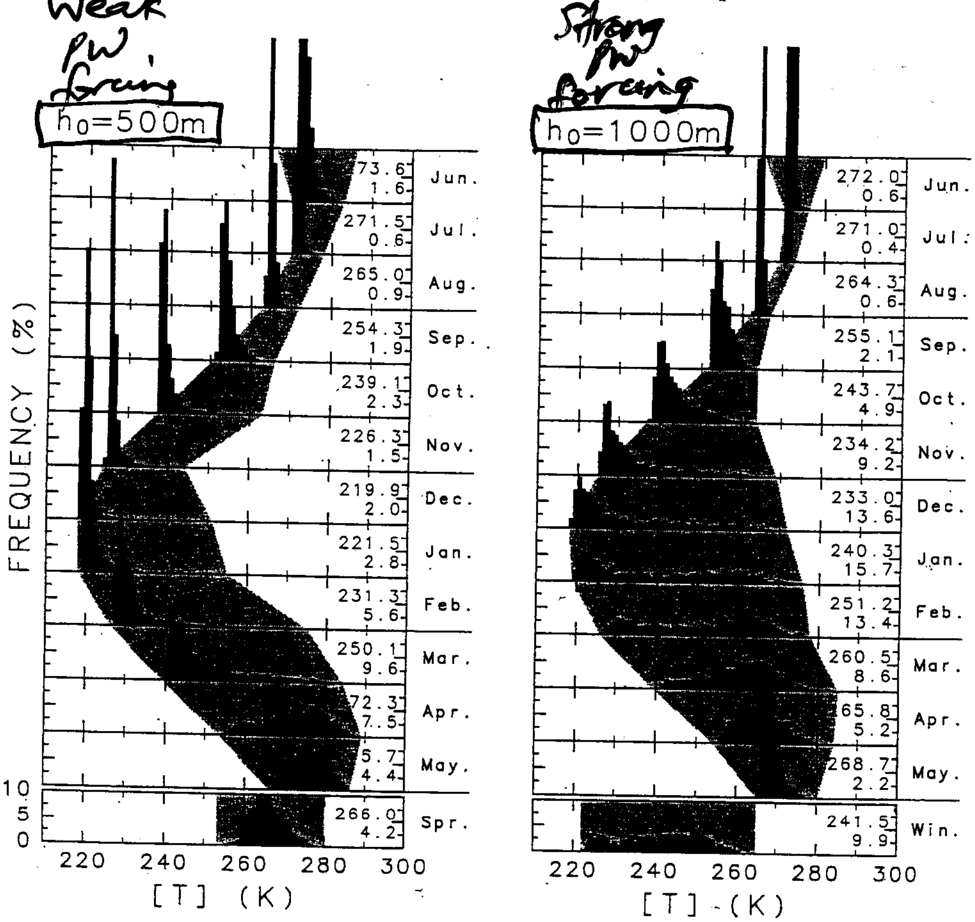
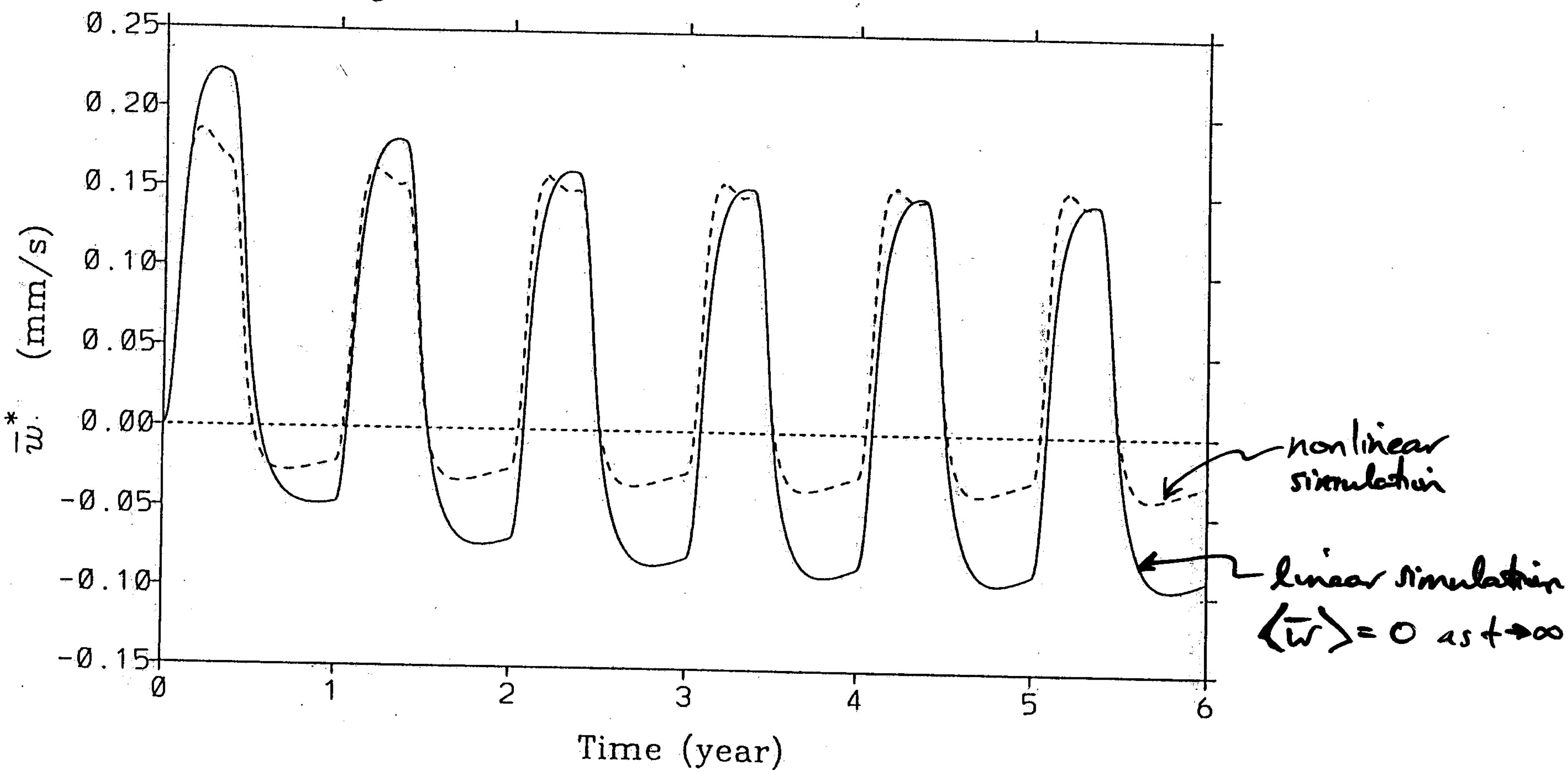


Fig. 10. Frequency distributions of the monthly mean temperature at 86°N and 2.6 hPa for the two millennium integrations: $h_0 = 500$ m (left) and 1000 m (right). Dashed line denotes the 1000-year mean annual variation of the monthly mean temperature, and shade shows the variable range. Averages and standard deviations for the 1000-year data are also written on the right hand side of each panel (top and bottom numbers, respectively). Frequency distributions for a seasonal mean are also displayed in the bottom: spring mean for $h_0 = 500$ m and winter mean for $h_0 = 1000$ m. The downward arrow in the seasonal mean indicates a threshold value for the 200 years of highest temperature.

Yoden, Taguchi & Naito (2002 JMSJ)

Upwelling at equator ($z = 33$ km) in a zonally symmetric balance model with time-periodic forcing imposed poleward of 15° latitude (outside Plumb & Eluszkiewicz equatorial viscous b.l.)

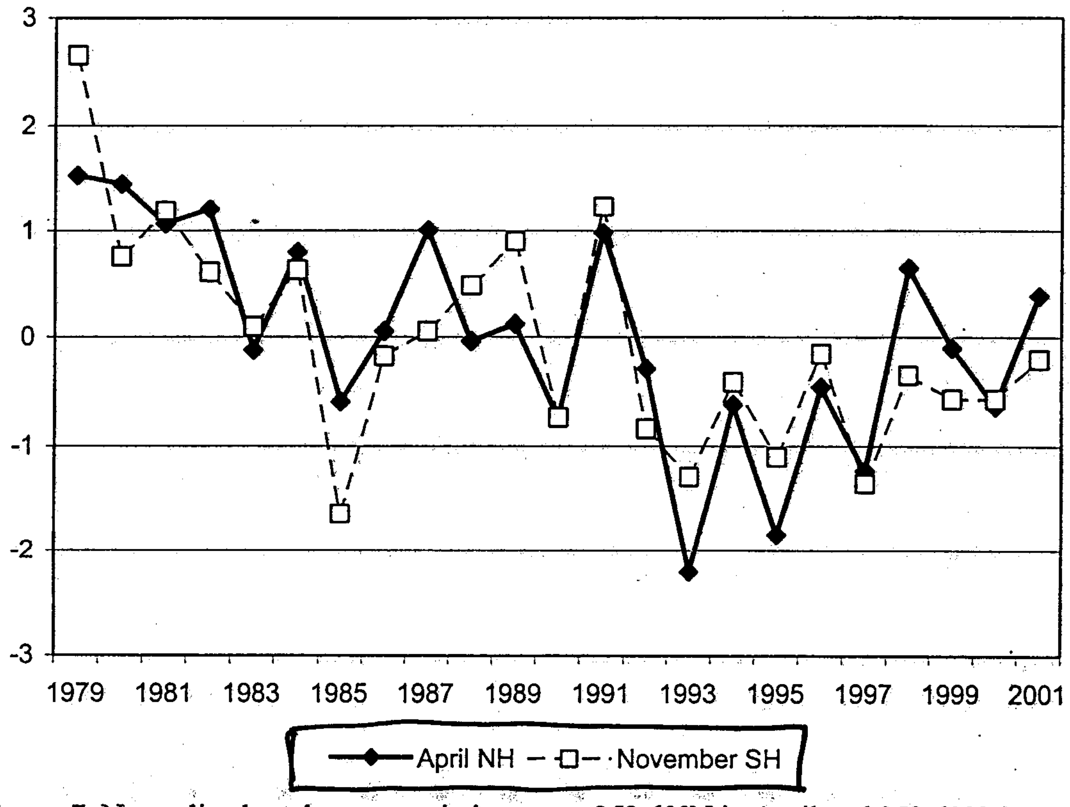
→ long transient!



Semeniuk & S.
JAS, ~~to appear~~ 2001

Calculations by Kirill Semeniuk, University of Toronto

April(NH) vs. following Nov. (SH)



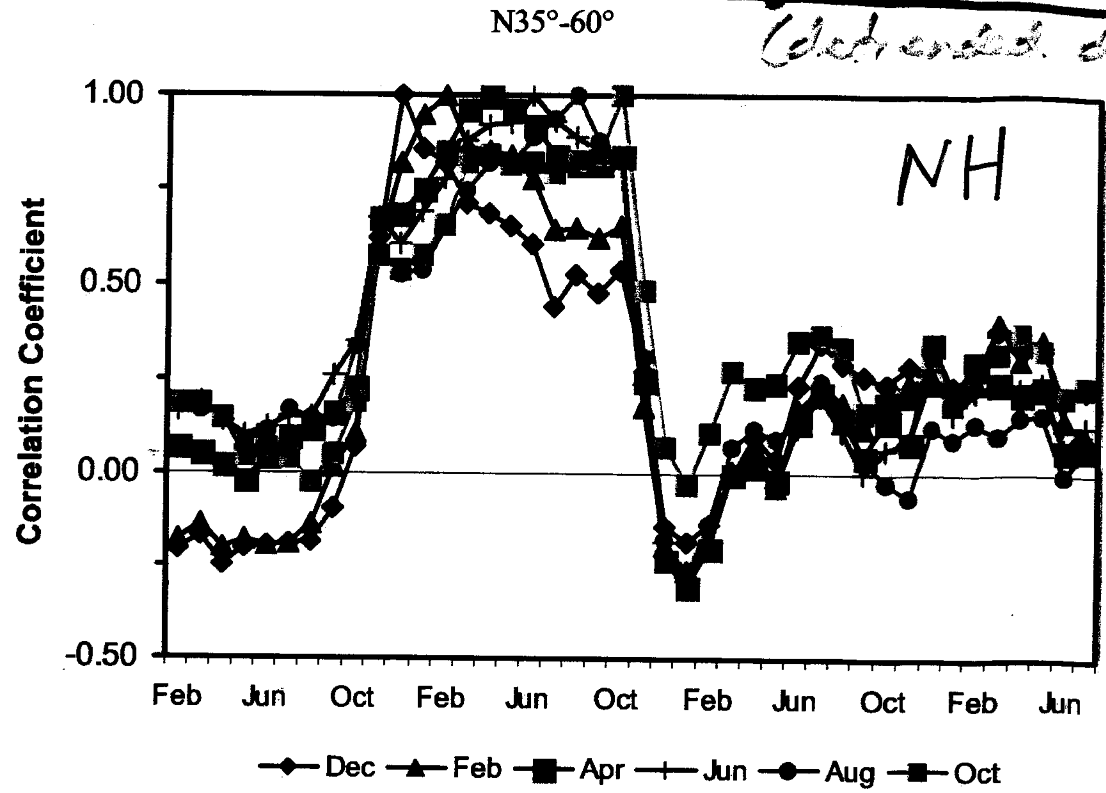
Fioletov & Shepherd (GRL 2003)

Figure 7. Normalized total ozone variations over 35°-60°N in April and 35°-60°S in November. The correlation coefficient between the two data sets is 0.81.

(mean = 0, std. dev = 1)

Calculations by Vitali Fioletov (Met. Service of Canada)

Autocorrelations of total ozone anomalies (detrended data)



Fioletov & Shepherd (GRL 2003)

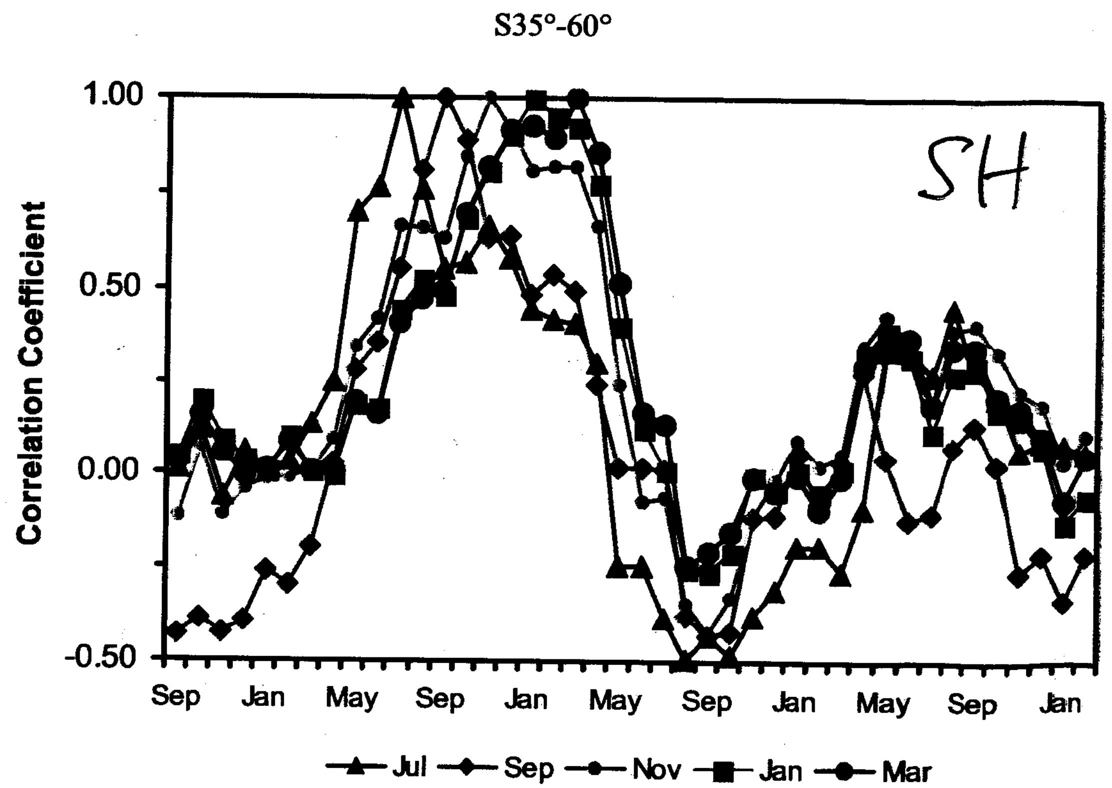
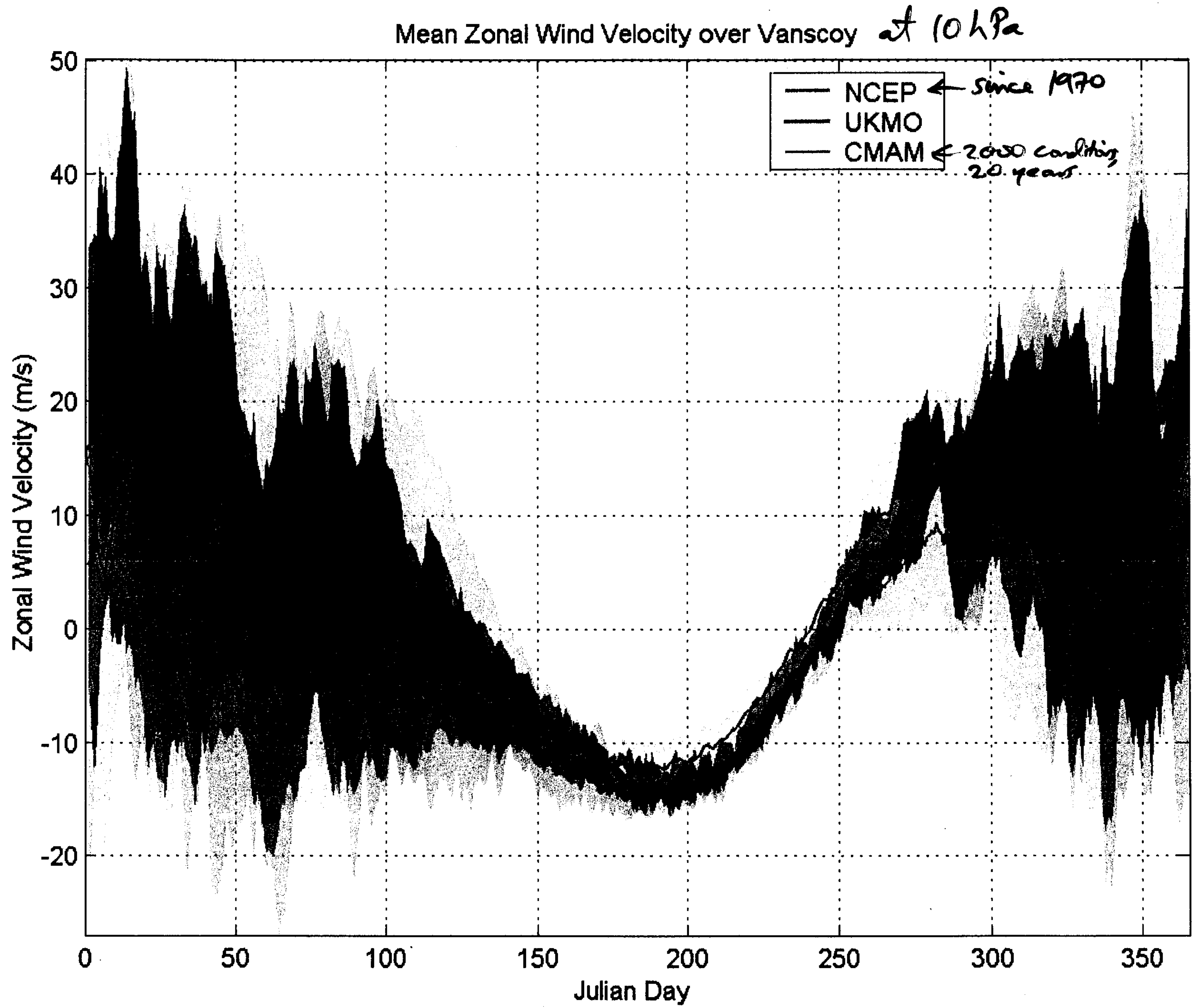


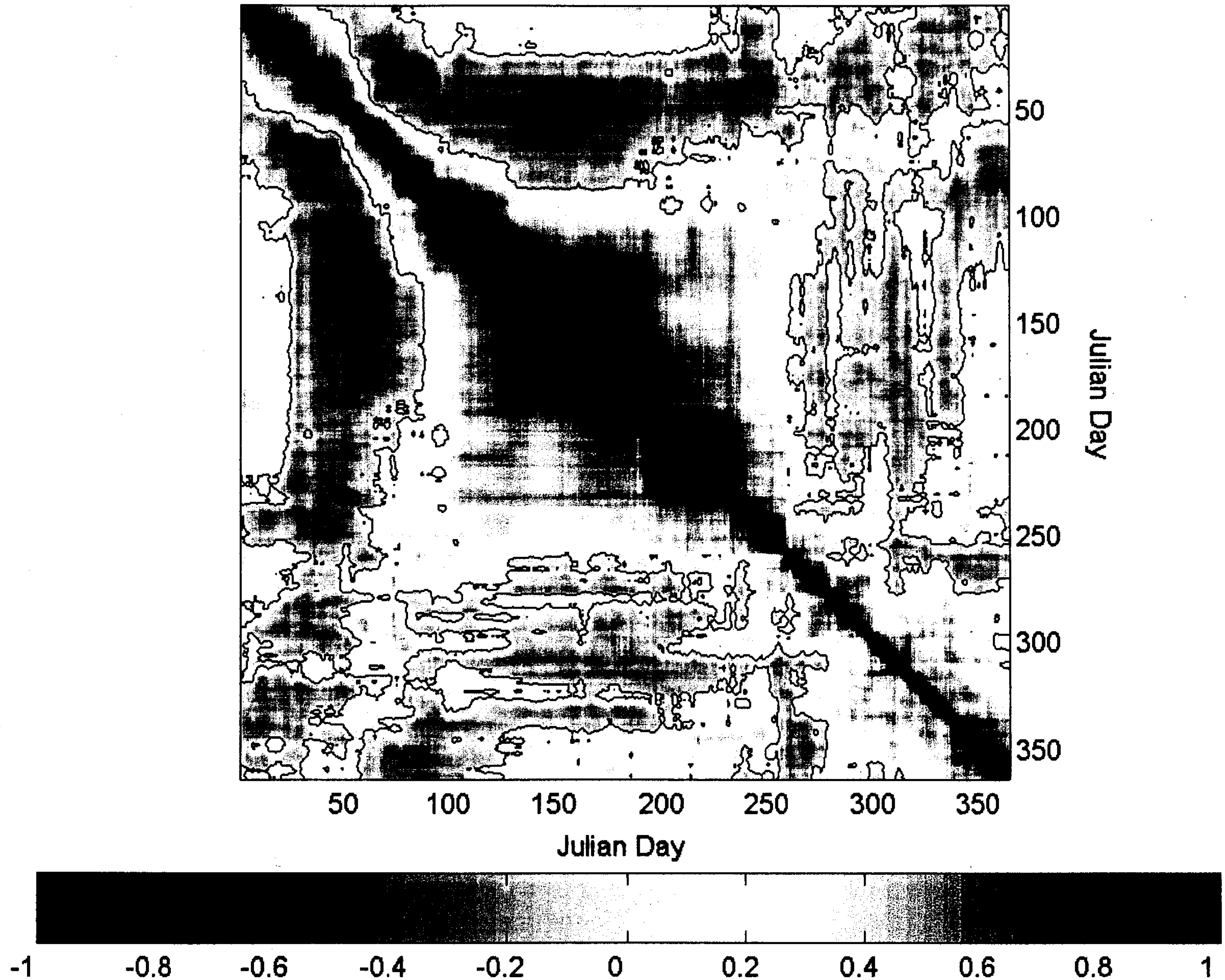
Figure 3. The same as Figure 2, but with monthly data detrended. A linear trend was estimated separately for each month of the year and then subtracted from the data.

Calculations by Vitali Fioletov (Met. Service of Canada)



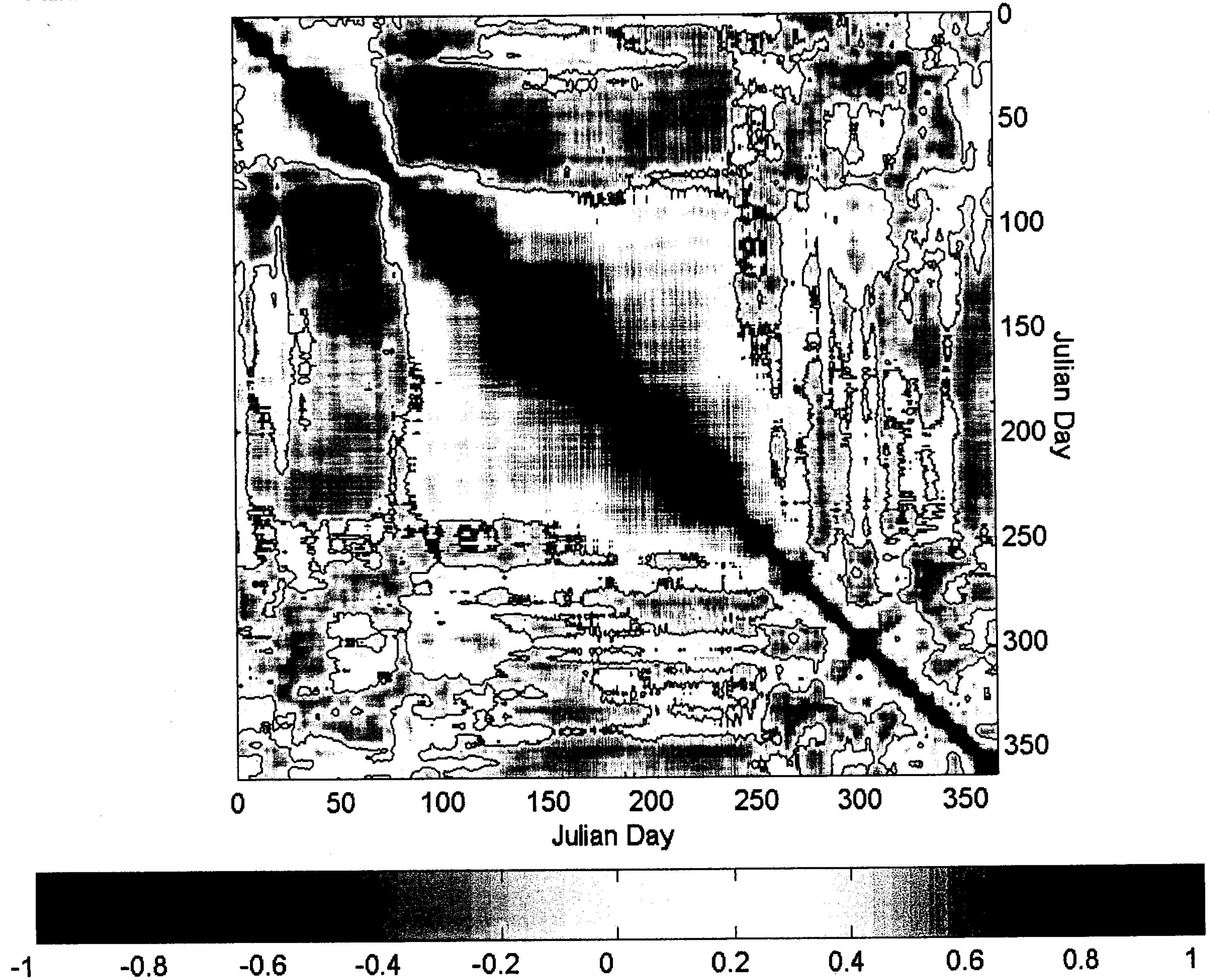
Courtesy of Debra Wunch (University of Toronto)

NCEP Autocorrelation Matrix for Mean Zonal Winds at 10 mb over Vanscoy



Courtesy of Debra Wunch (University of Toronto)

Autocorrelation Coefficients for Mean Zonal Winds at 50.625°N using 24 years of CMAM Data



Courtesy of Debra Wunch (University of Toronto)

Seasonality of regression coefficients with April (NH) and November (SH)

Fioletov & Shepherd (GRL 2003)

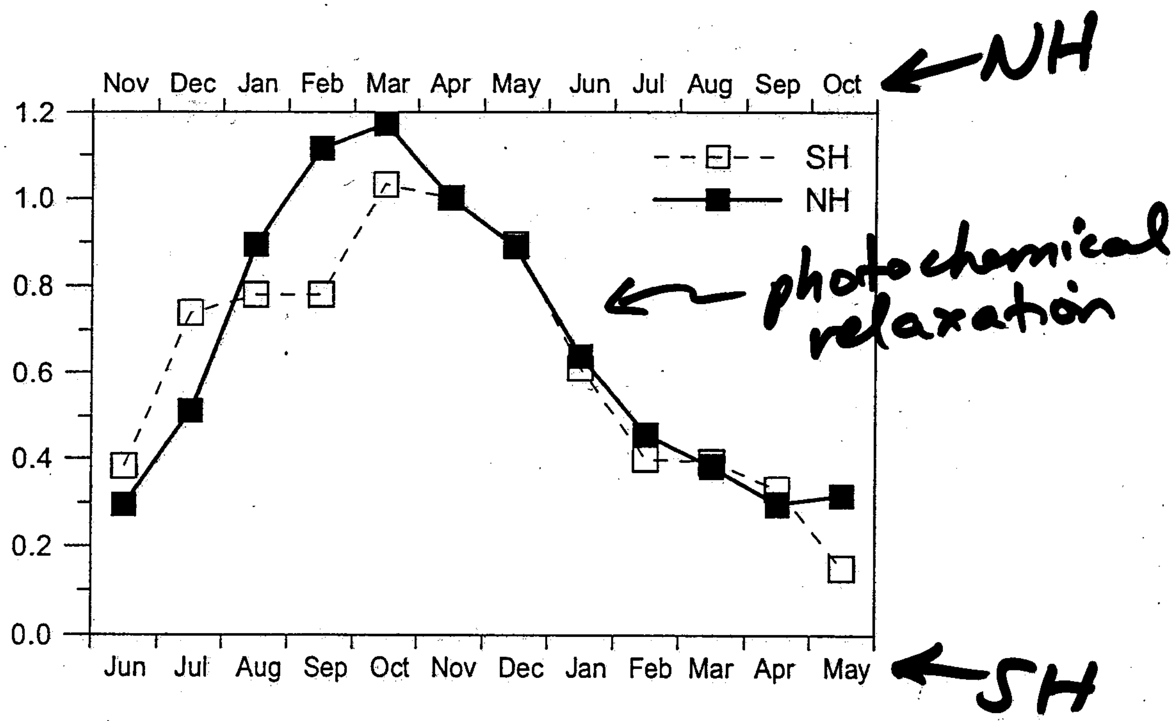


Figure 5. Linear regression coefficients (solid line with black squares) between ozone departures for April and ozone departures in other month of the year (shown on the top) for the 35°-60°N zone. Similarly, dashed line with open squares shows regression coefficients between November departures and ozone anomalies in other months for the 35°-60°S zone. Detrended data were used.

Calculations by Vitali Fioletov (Met. Service of Canada)

Trends estimated by regression from other months

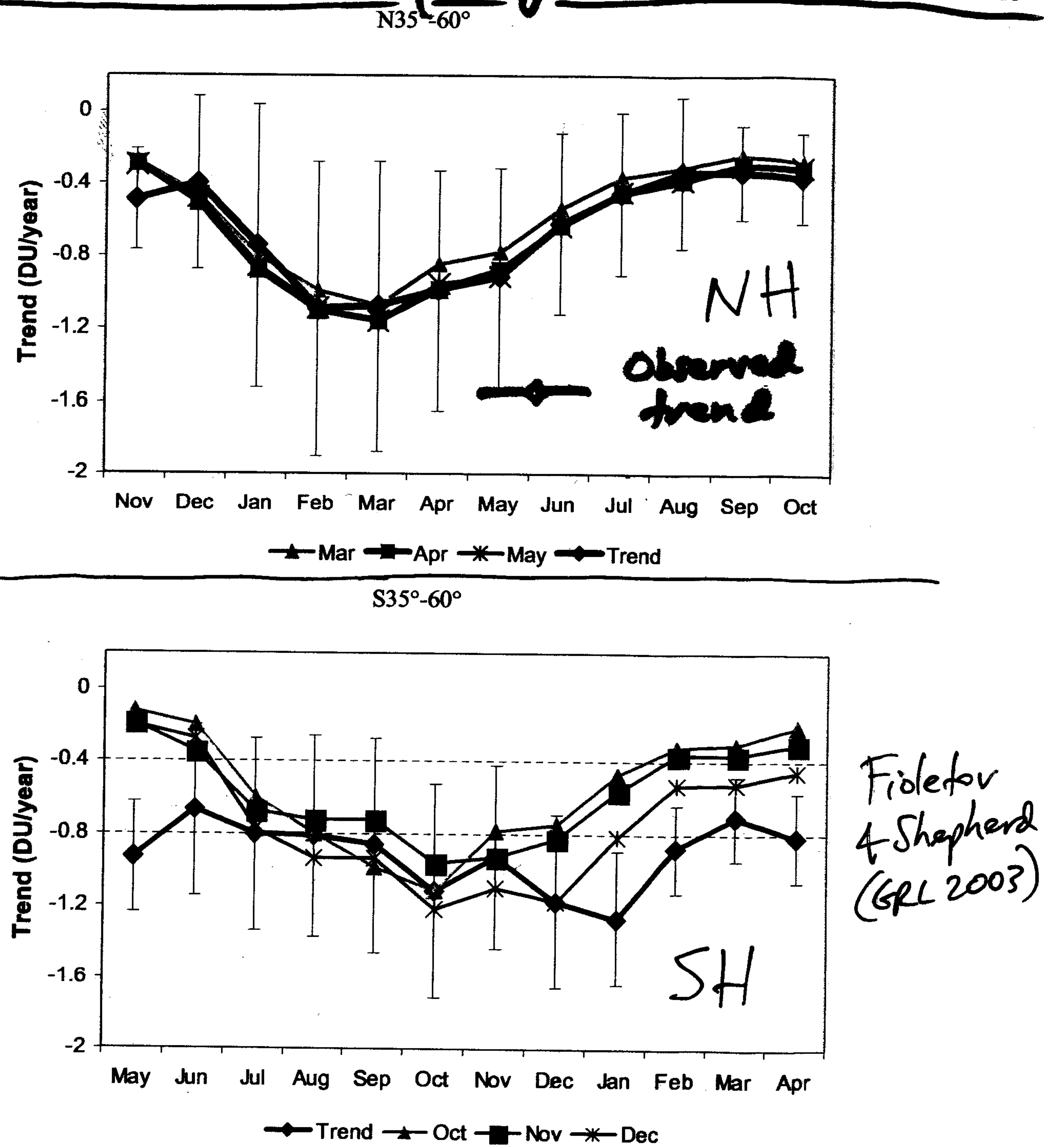
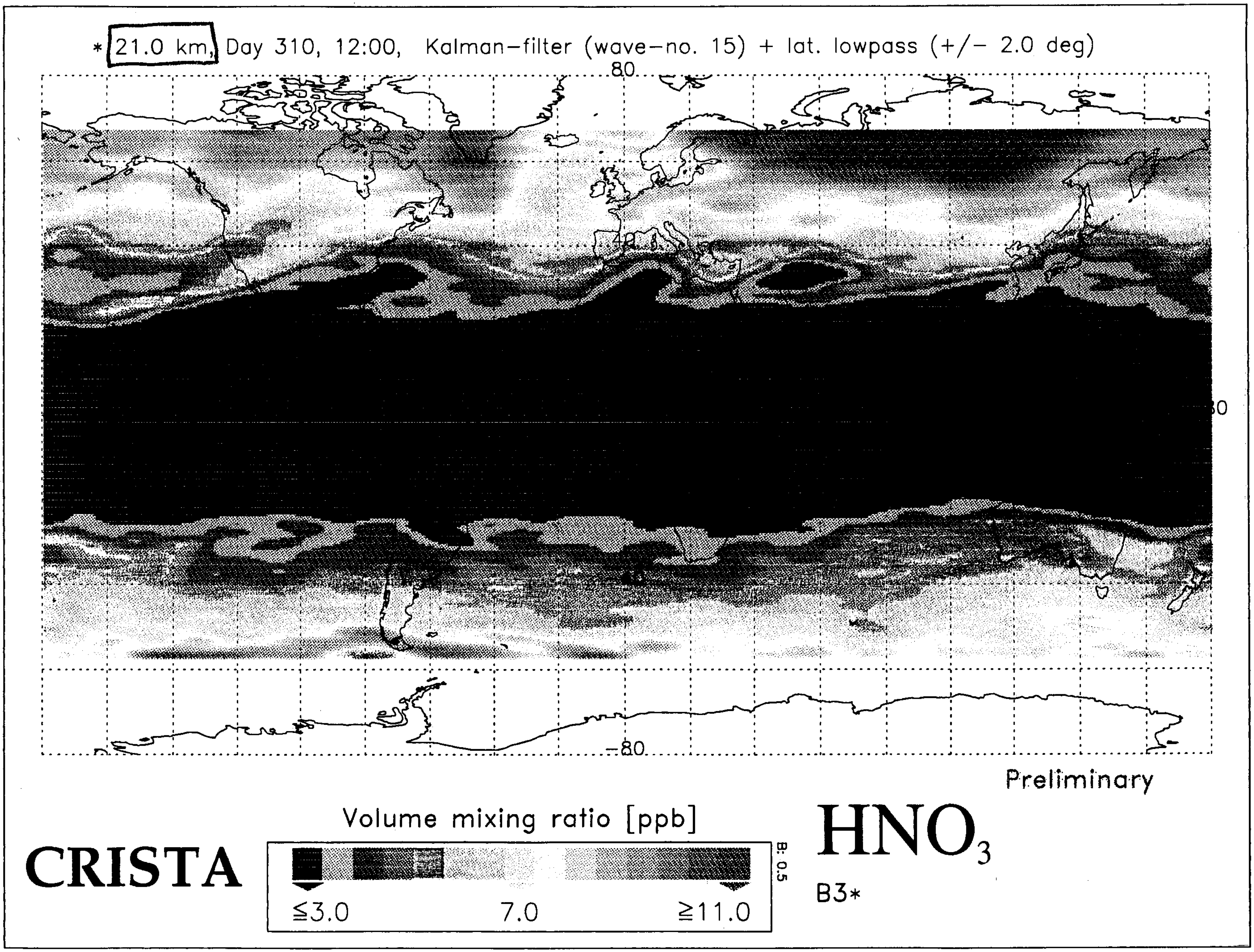


Figure 5. (top) Total ozone trends at 35°-60°N in different months of the year for the 1979-2001 period with 95% confidence limits (solid black line). Total ozone trends estimated from a linear trend in March (green), April (red), and May (magenta) and from a linear regression coefficient between March (April, May) detrended values and values in other months. (bottom) Similar plot for 35°-60°S.

Calculations by Vitali Fioletov
(Met. Service of Canada)

Courtesy of D. Offermann (Wuppertal Univ, Germany)



May 26 14:13 1997 /d2/filter/kalman/ein/fillp/b3_hno3_r2n_310_21_w15_2.fillp

17a

Garcia & Boville (1994, J. Atmos. Sci.)

Zonally symmetric model with parameterized planetary and GW drag

——— with GWD
– – – – without GWD

→ Sensitivity is strong in the polar night, due to long radiative relaxation times

Southern Hemisphere wintertime

Northern Hemisphere wintertime

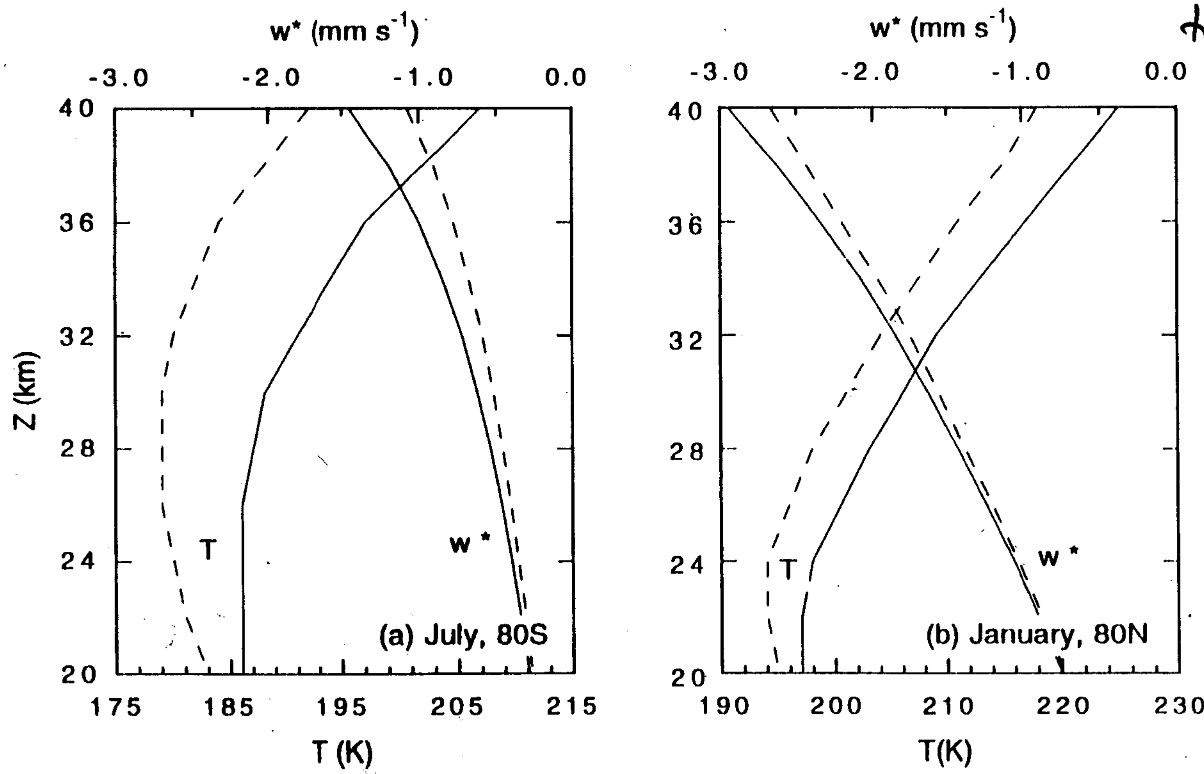
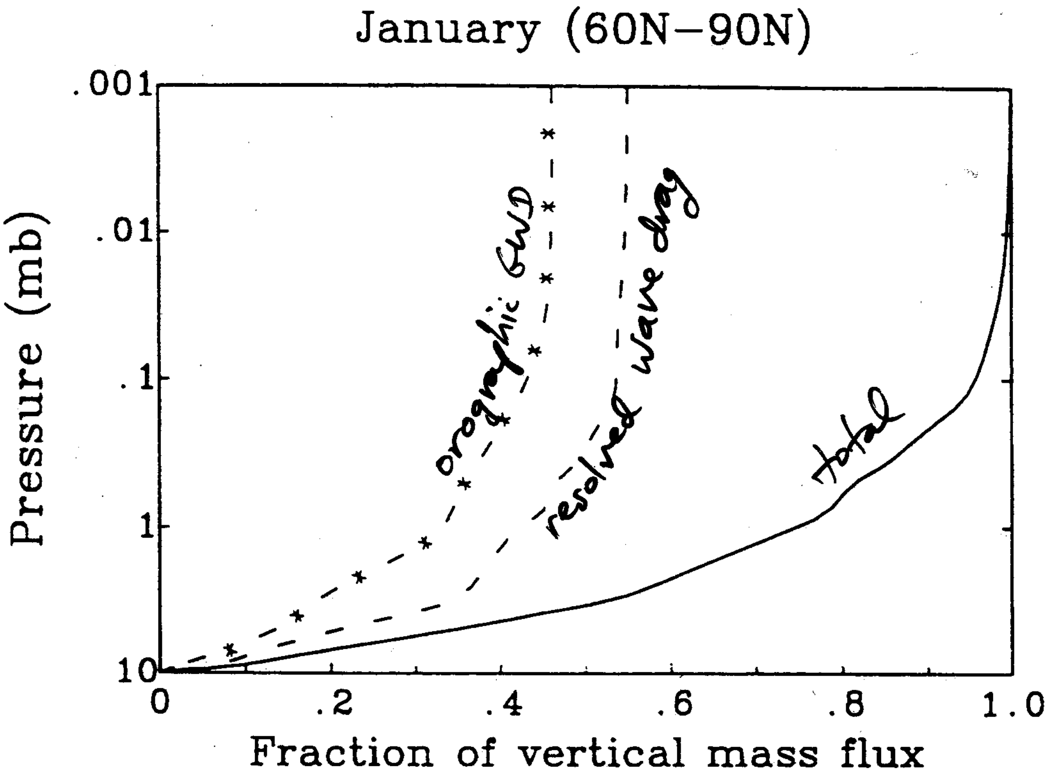


FIG. 4. Vertical profiles of mean vertical velocity and temperature for (a) July at 80°S and (b) January at 80°N obtained when the model is run with parameterized gravity wave driving (solid lines) and without it (dashed lines).

Forcing of downwelling over polar cap at 10 mb in CMAM

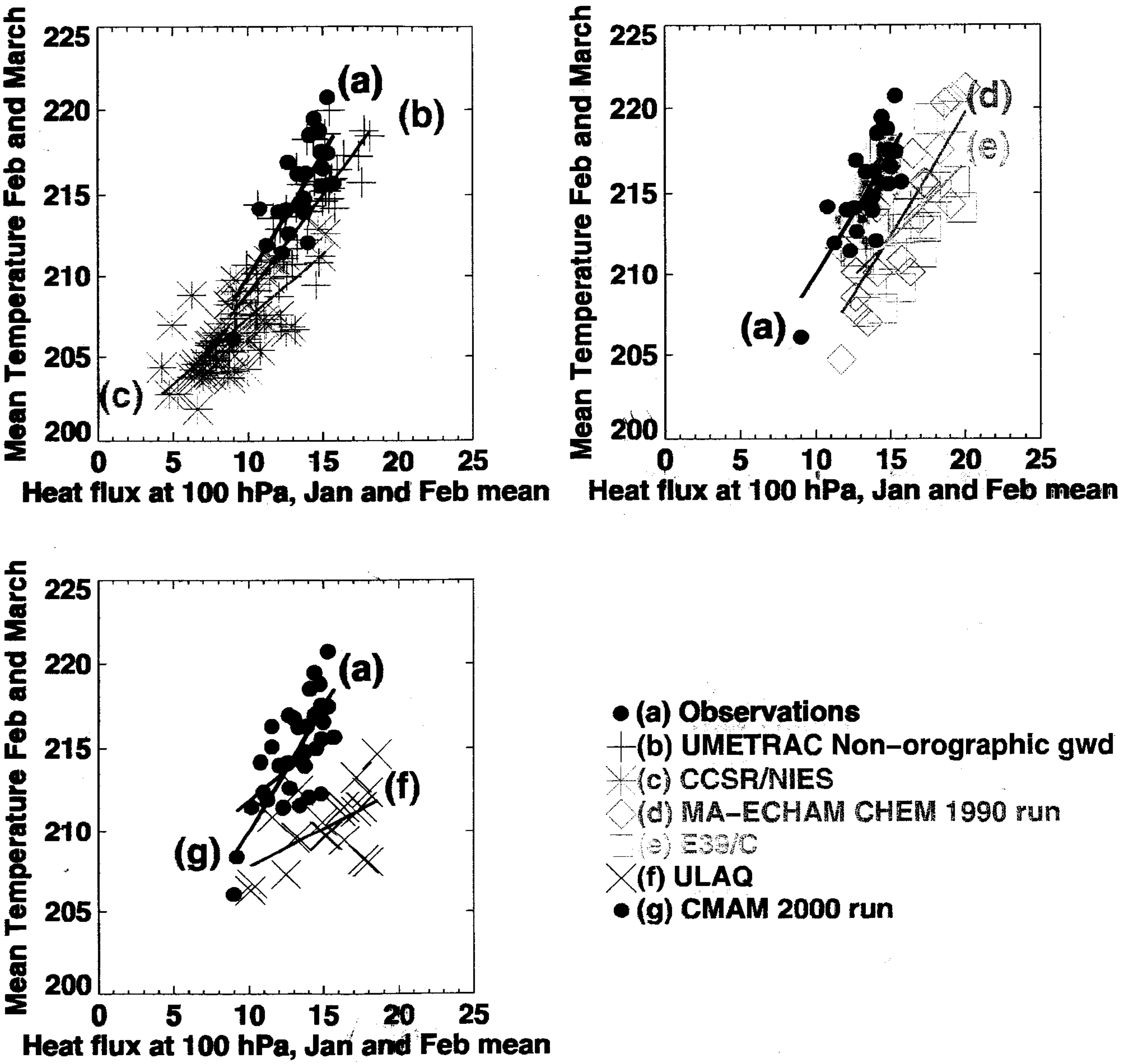


Downward control : $\rho_0 \overline{w}^*(z) = \int_z^\infty \frac{\partial}{\partial y}\left(\frac{\rho_0 F}{f}\right) dz'$, F = wave drag

- GWD is a large part of the forcing in NH winter
- Much of the drag is located well above the altitude of interest

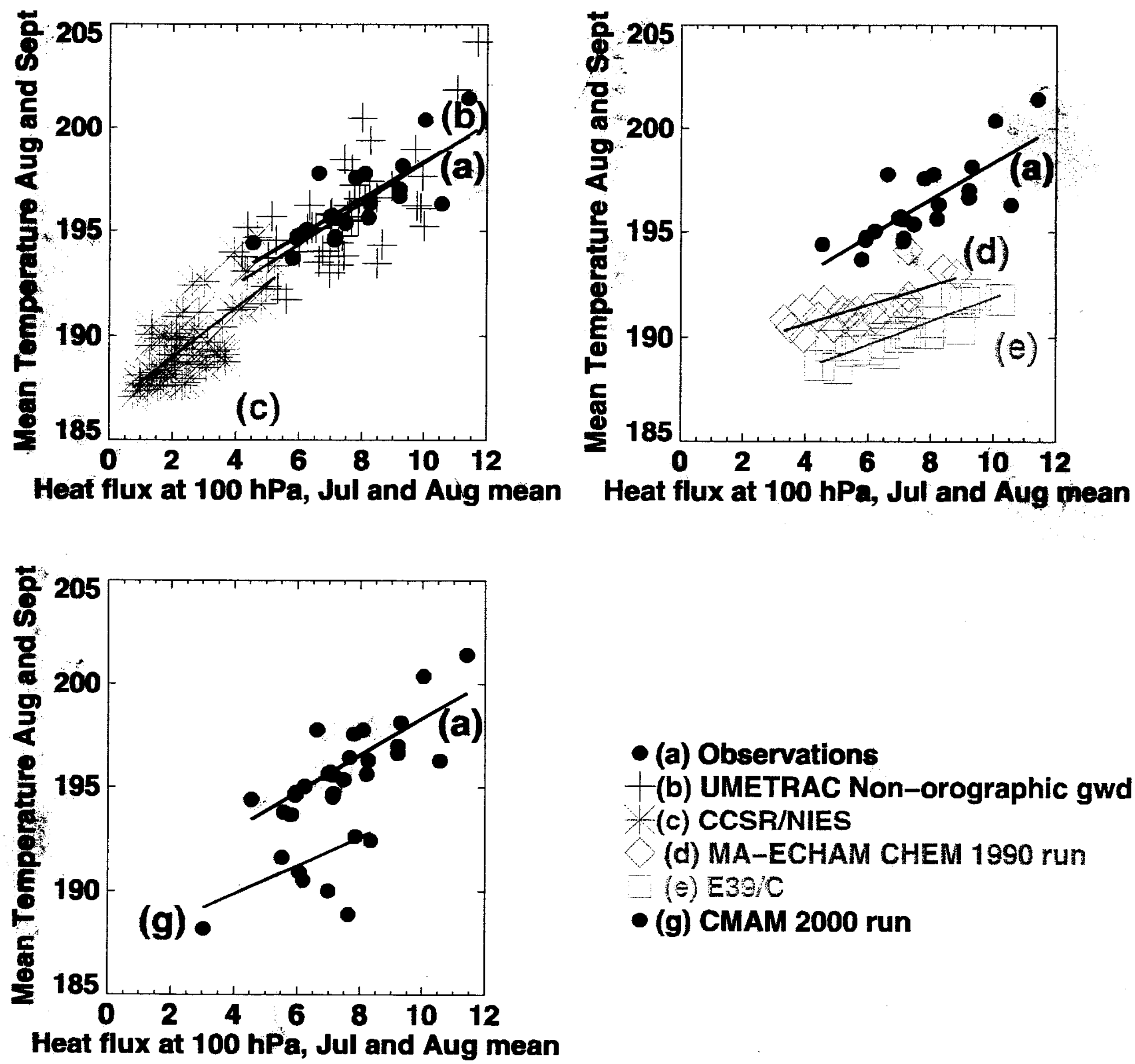
from Beagley et al. (1997, Atmos-Ocean)

Relation between stratosphere wave driving and polar T (NH)



Austin et al. (2003 ACP)

Relation between stratospheric wave drag and polar T (SH)



Austin et al. (2002 ACP)

Why does momentum conservation matter?

For gravity waves, in steady state,

$$\bar{w}^* = \frac{-1}{a\cos\phi}\frac{\partial}{\partial\phi}\left\{\frac{\overline{u'w'}\cos\phi}{2\Omega\sin\phi}\right\} \qquad \text{(Haynes et al. 1991 JAS)}$$

⇒ Downwelling depends on net drag above the level in question, not on where the drag is exerted

⇒ Changes in winds above cannot change $\bar{w}^*$

However this is not true for Rayleigh drag, or for non-conservative GWD schemes

⇒ Such schemes could lead to spurious downward influence

N.B. Taking away the drag affects $\bar{w}^*$; but this is not a realistic thought expt!

Zonally symmetric response to imposed force within the sponge layer

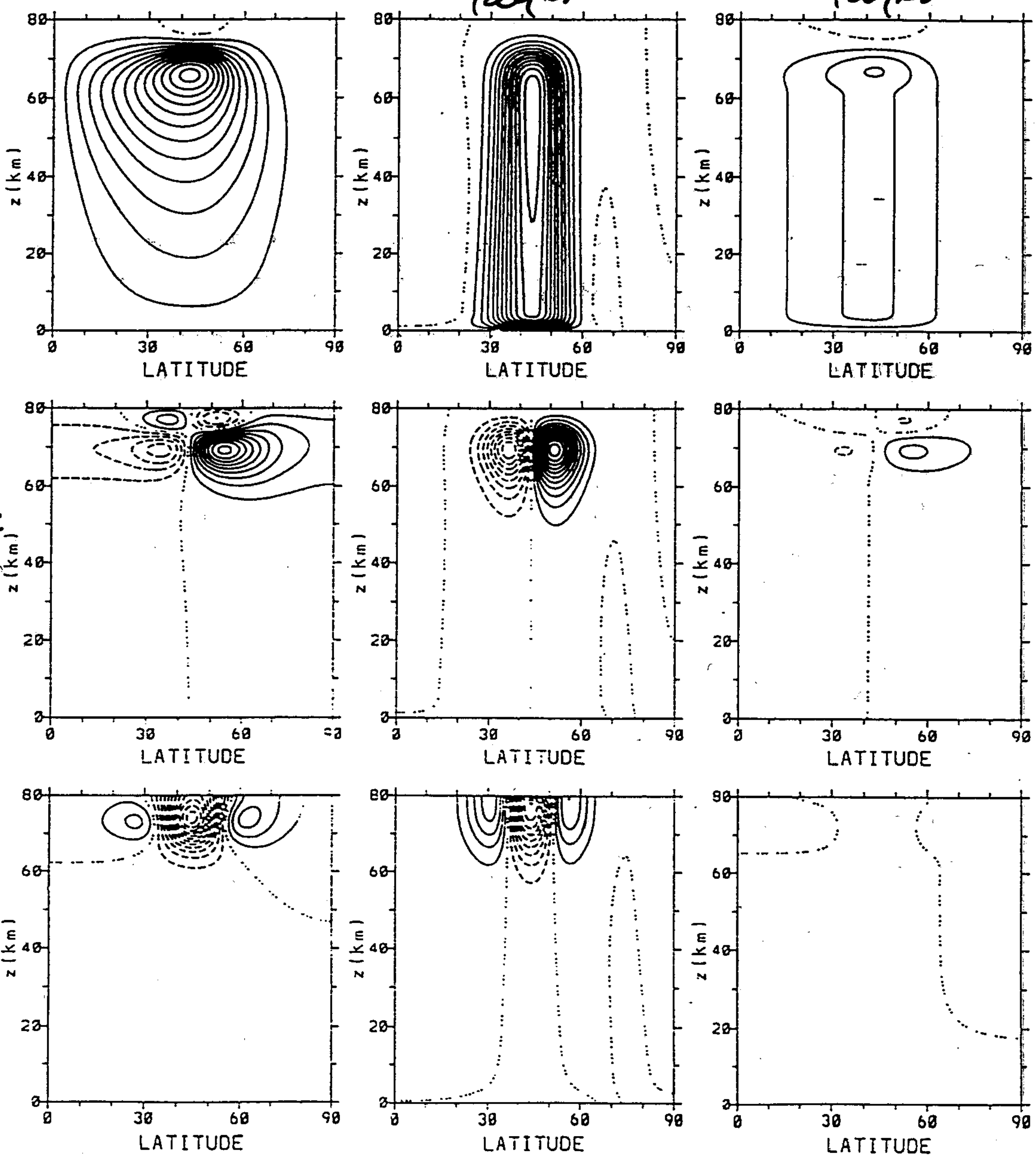
→ shows futility of applying GWD within a sponge layer: implies model lid ≥ 110 km if mesopause drag is needed and sponge is retained

Instantaneous (a) | Steady, no sponge layer (b) | Steady, with sponge layer (c)

Residual ψ Circulation (i)

Temperature T (ii)

Zonal wind u (iii)

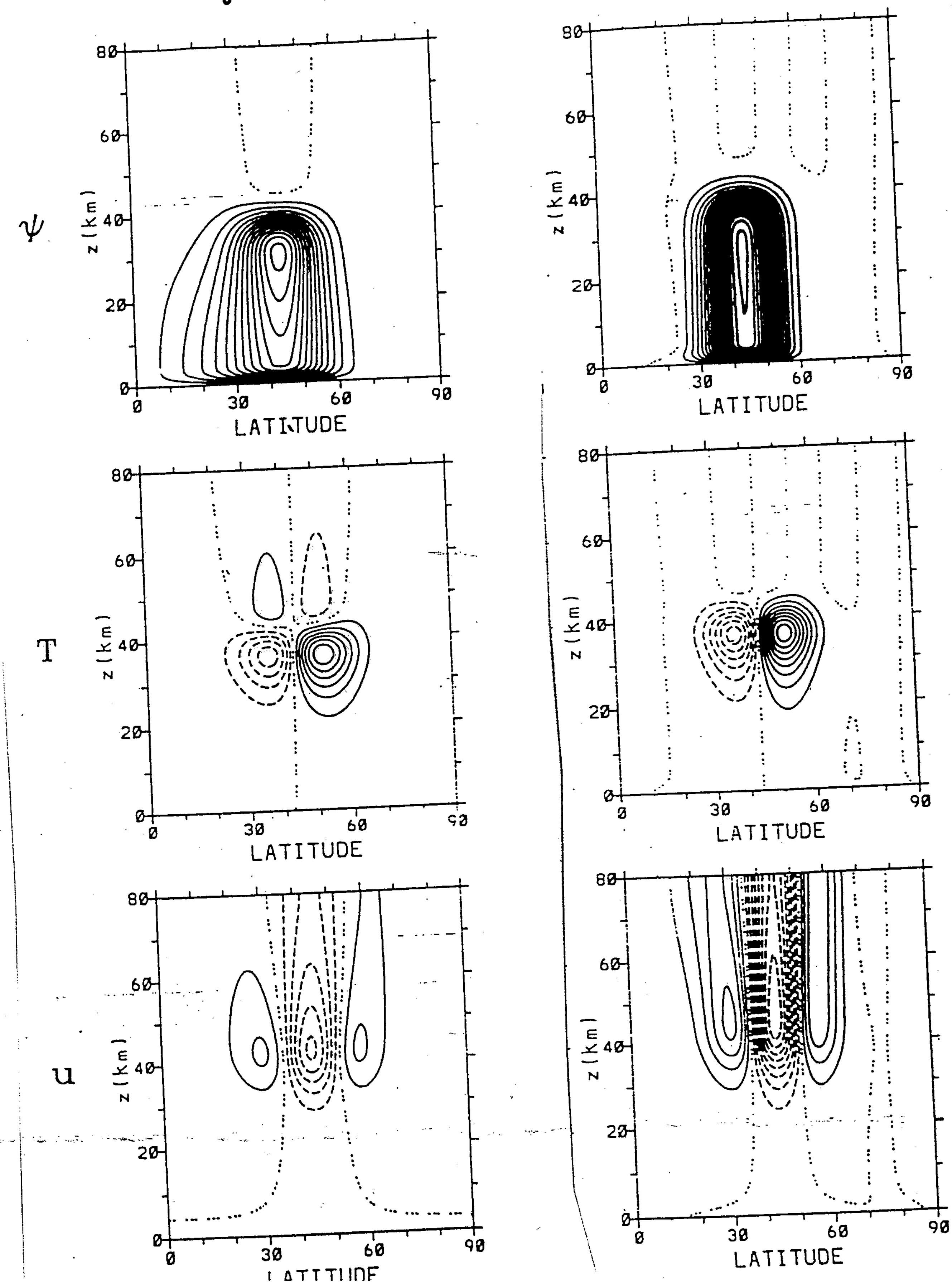


Calculations by Kirill Semeniuk, Univ. of Toronto

Shepherd, Semeniuk & Koshyk (JGR, 1996)

Response to imposed forcing ($\alpha = 1/5$ days)

Uniform Rayleigh drag ($\tau = 1/100$ days) — only b.l. drag



Time evolution of stratospheric wave driving

Models forced by changing halogens and GHGs

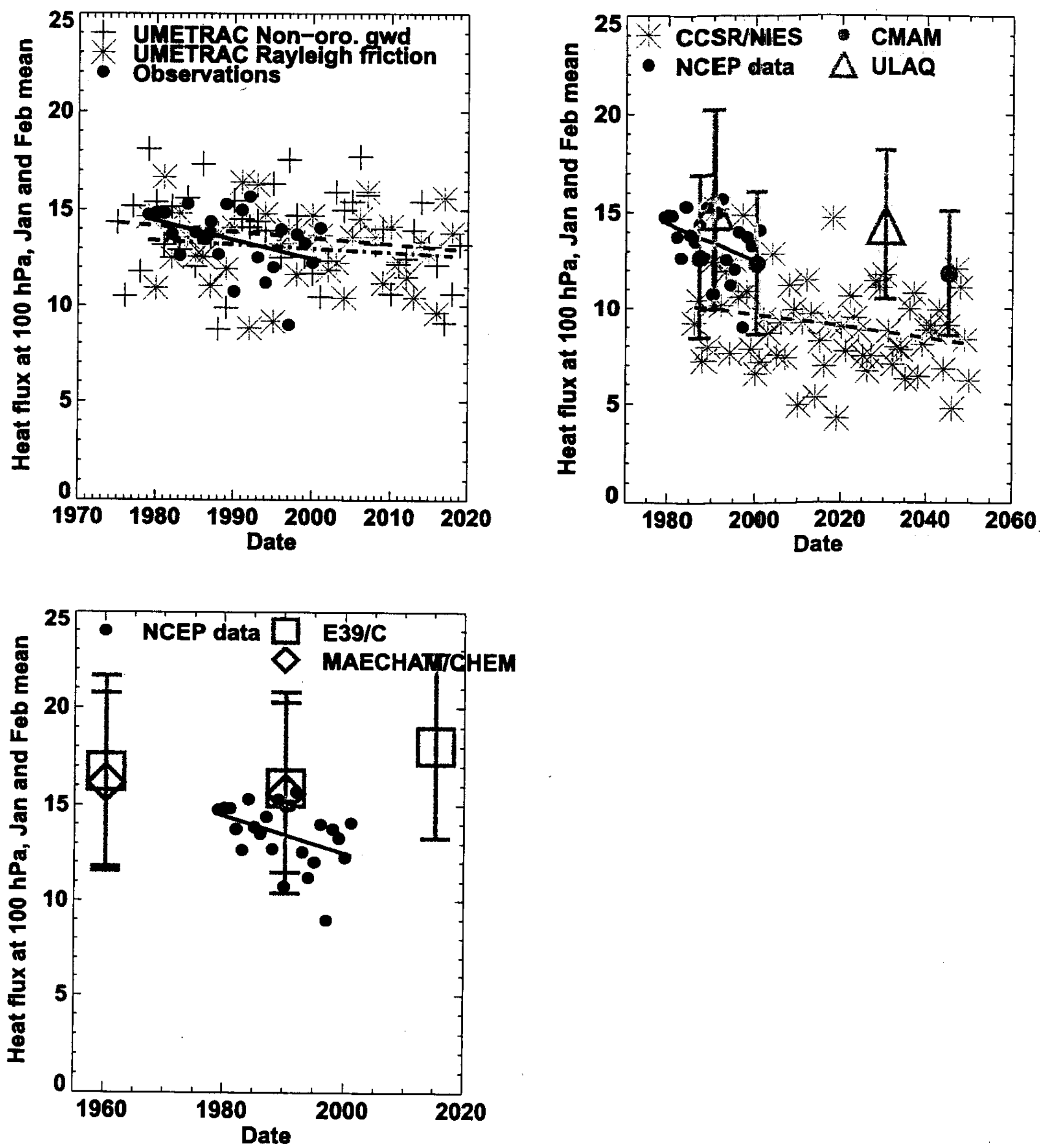


Fig. 8. Scatter diagrams of heat flux $\overline{v'T'}$ (averaged 40°–80° N, at 100 hPa for January and February) against year for participating models. In all panels, the linear regression line between the NCEP derived heat flux and time is drawn as a solid line. Upper left panel: the dashed line is the linear regression for the non-orographic gwd run of UMETRAC, and the dot-dash line is the linear regression for the Rayleigh friction run of UMETRAC. Upper right panel: the dashed line is the linear regression for the CCSR/NIES results. Two standard deviations of the annual values are indicated by the error bars for the time slice experiments. For CMAM, the results are plotted for 2045 rather than 2028 since the results are dependent largely on the WMGHG concentrations (see Sect. 2.1).

Austin et al. (ACP 2003)

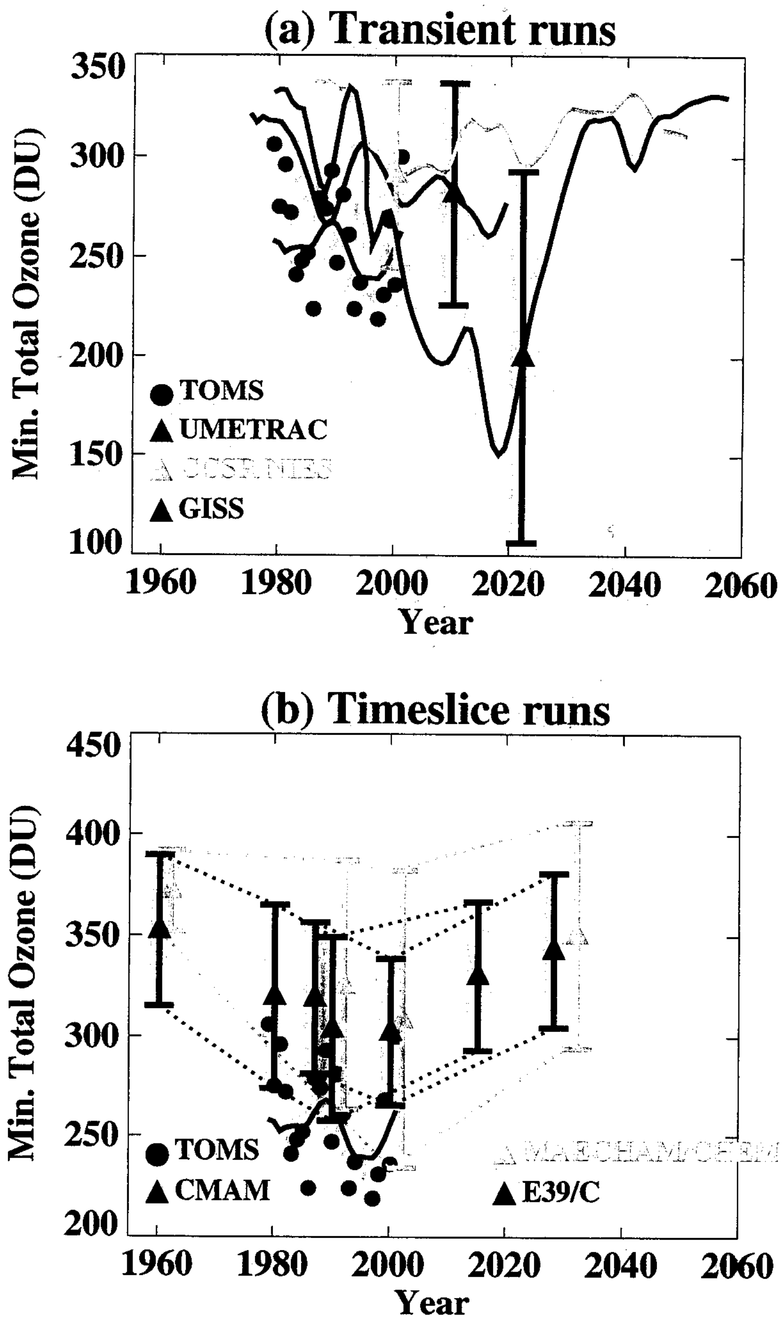


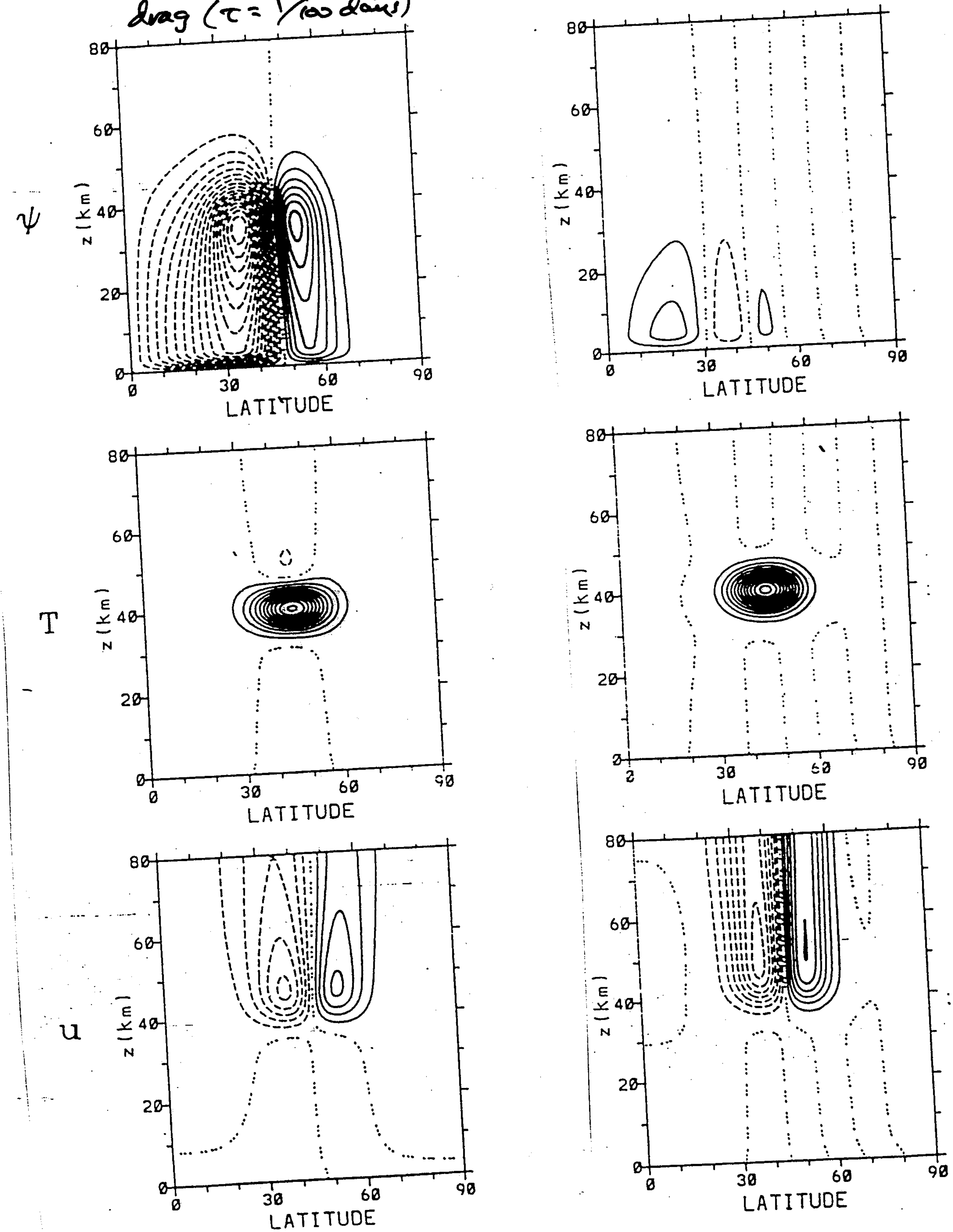
Fig. 9. Minimum Arctic (March/April) total ozone for the main experiments of this assessment. **(a)** Transient runs in comparison with TOMS data. The solid lines show the results of a guassian smoother applied to the individual year's results. The error bars denote twice the standard deviation of the individual years from the smoothed curve. **(b)** Time slice runs in comparison with TOMS data. The error bars denote the mean and twice the standard deviation of the individual years within each model sample (10 years for CMAM, 20 years for MAECHAM/CHEM and E39/C). Dotted lines are drawn between the end points of the error bars to assist in estimating trends by eye. For MAECHAM/CHEM only: (i) the values have been plotted two years late for clarity, (ii) a standard tropospheric column of 100 DU has been added to the computed columns above 90 hPa. Note that the MAECHAM/CHEM results are not symmetric about the mean, but have a long tail towards low values.

Austin et al. (ACP 2003)

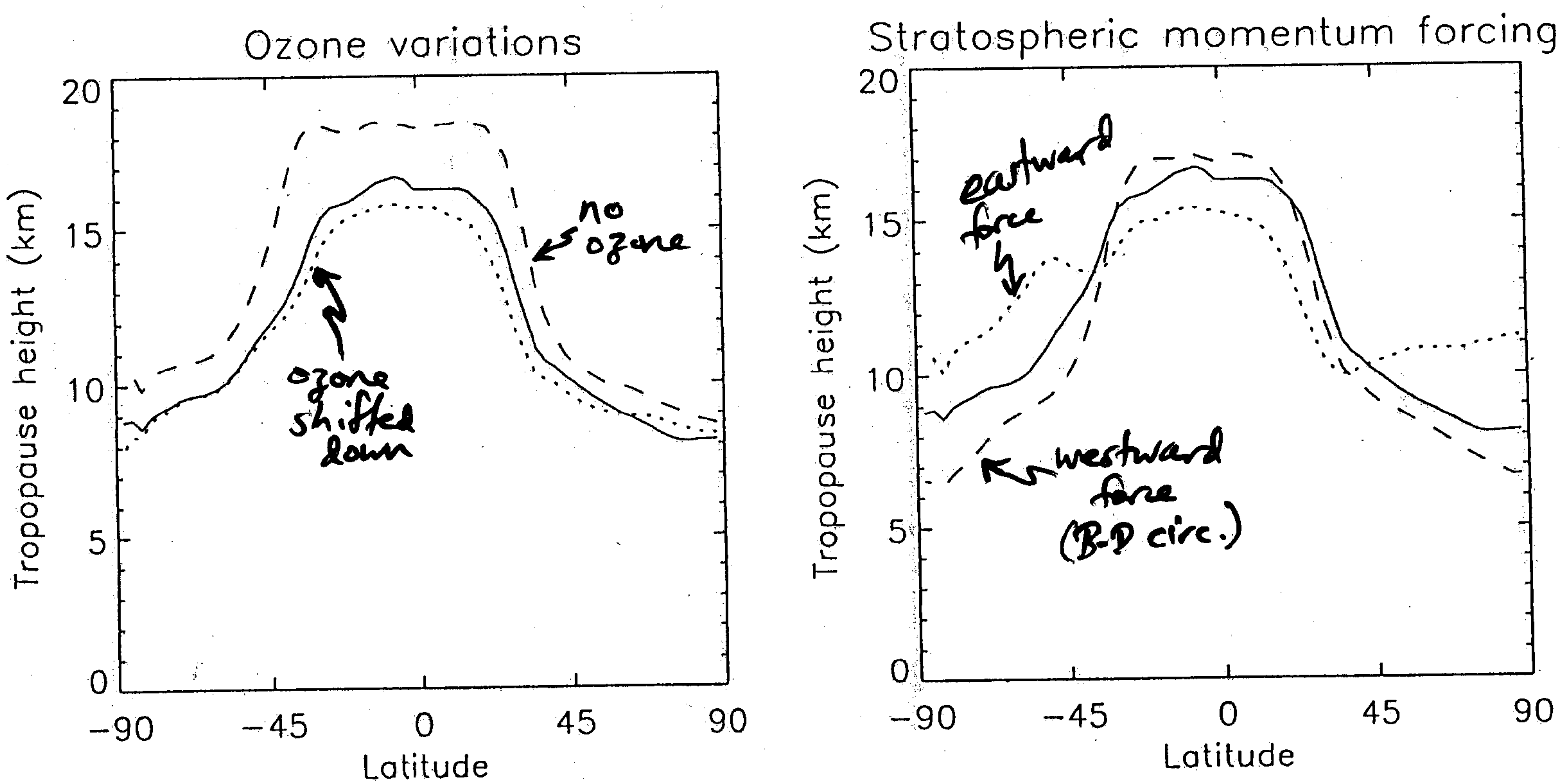
Response to imposed heating (…)

Uniform Rayleigh drag (τ = 1/100 days)

Only b.l. drag



Zonal-mean tropopause height (lapse-rate criterion) for GCM experiments with changes in ozone or prescribed stratospheric forces



Thuburn + Craig (2000, JAS)

Stationary-wave geopotential height

Observations

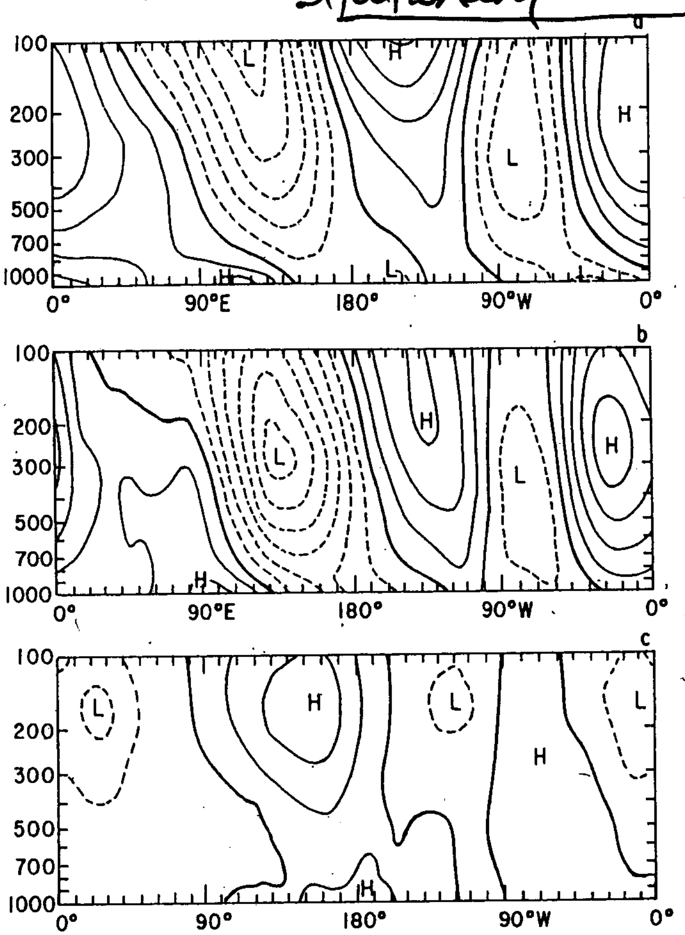


Fig. 2.3. Longitude–height cross-sections of stationary-wave geopotential height $\bar{Z}^*$ for the winter season, derived from 11 years of NMC operational analyses, adapted from Lau (1979). (a) 60°N; (b) 45°N; (c) 25°N. Contour interval 50 m; the zero contour is thickened.

from Wallace (1983)

Linear theory

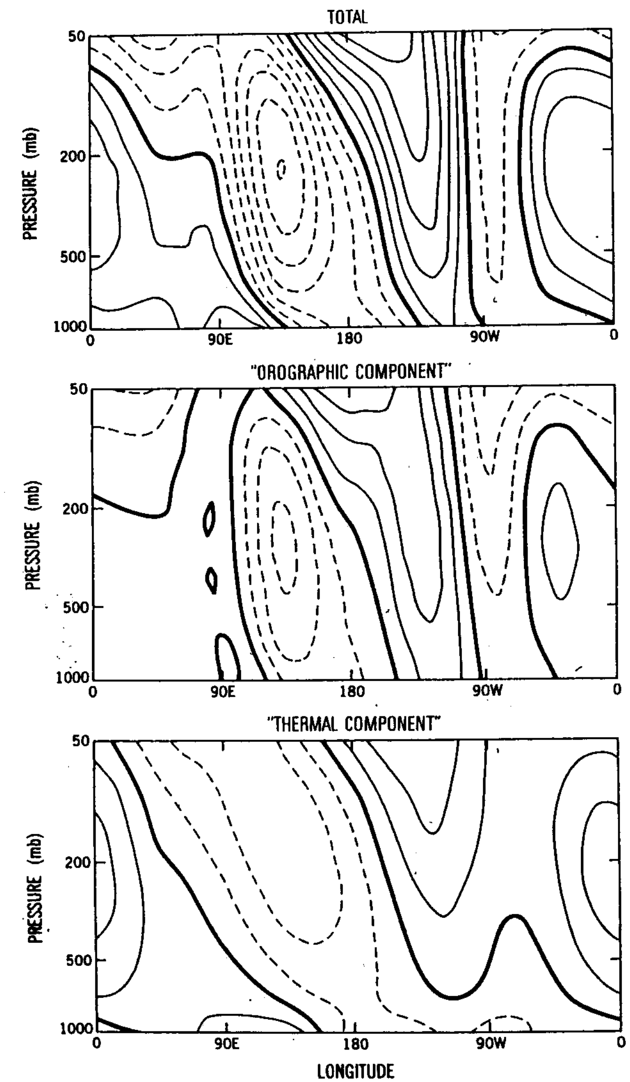
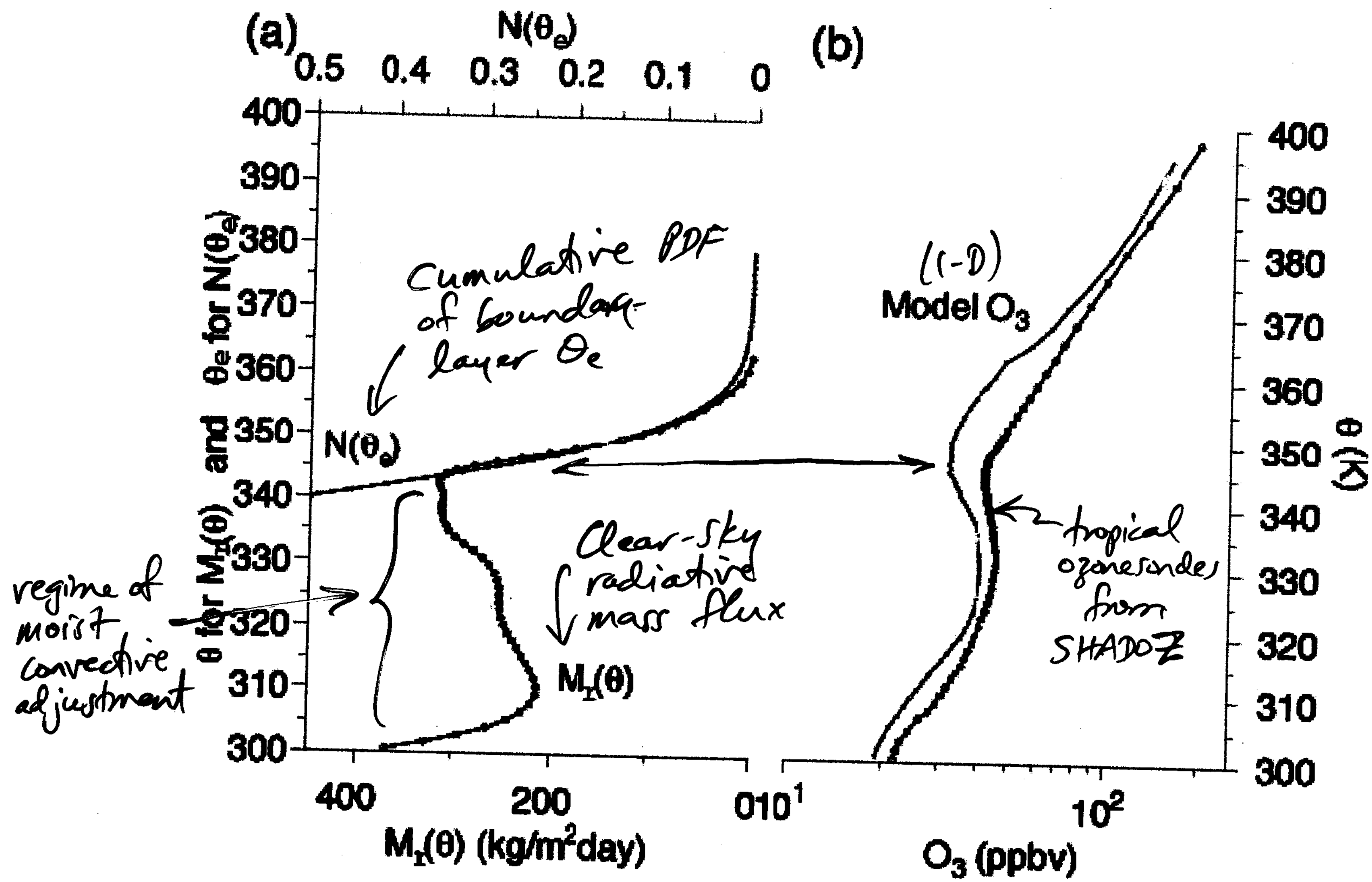


Fig. 6.20. The eddy geopotential height as a function of longitude and pressure at 45°N for the mountain model ('total'), the mountain minus no-mountain flow ('orographic component') and the no-mountain model ('thermal component'). The contour interval is 50 m with negative contours dashed. Model fields in Figs. 6.19–6.23 are averaged over the three winter months in each of the 15 model years.

from Held (1983)

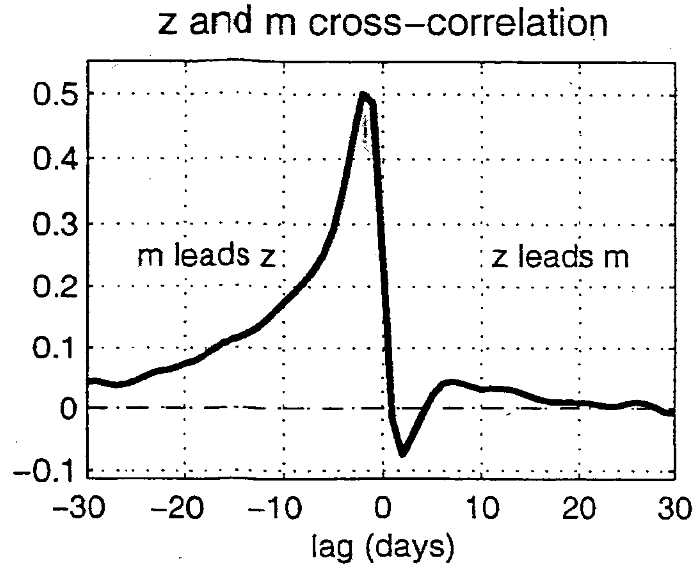
In "Large-Scale Dynamical Processes in the Atmosphere" (ed. Hoskins & Pearce), 1983, Academic Press

Mass-flux scaling in the tropical tropopause layer



Courtesy of Ian Folkins (Dalhousie Univ.)
(Folkins 2002 JAS)

Cross correlation between SAM projection of $\bar{u}$ (z) and of the eddy momentum flux convergence (m) (vertically averaged)

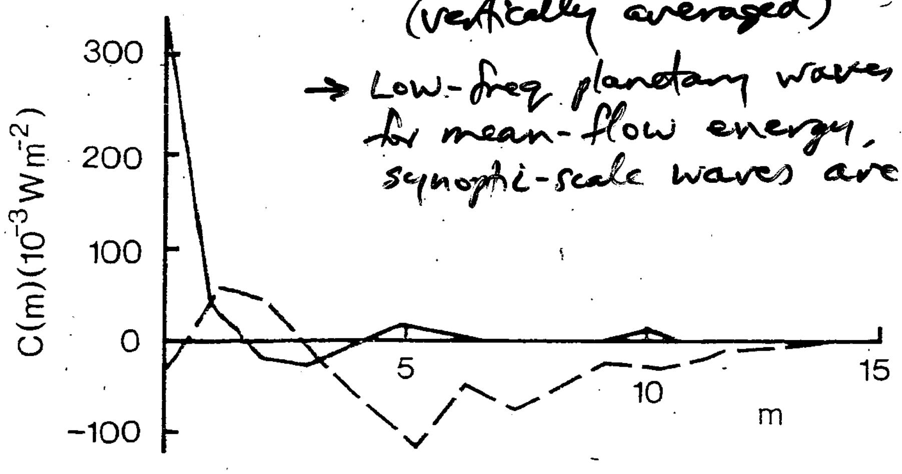


→ Changes in $\bar{u}$ induced by changes in m

Lorenz & Hartmann (2001 JAS)

FIG. 5. Cross correlation between z and m.

Energy conversion terms between stationary flow (solid) and transient flow (dashed), as a function of zonal wavenumber m (vertically averaged)



→ Low-freq. planetary waves are a sink for mean-flow energy, while synoptic-scale waves are a source

FIG. 8. Energy conversion terms $C_S(m)$ (solid) and $C_T(m)$ (dashed), as functions of zonal wavenumber m, shown only up to $m \leqslant 15$.

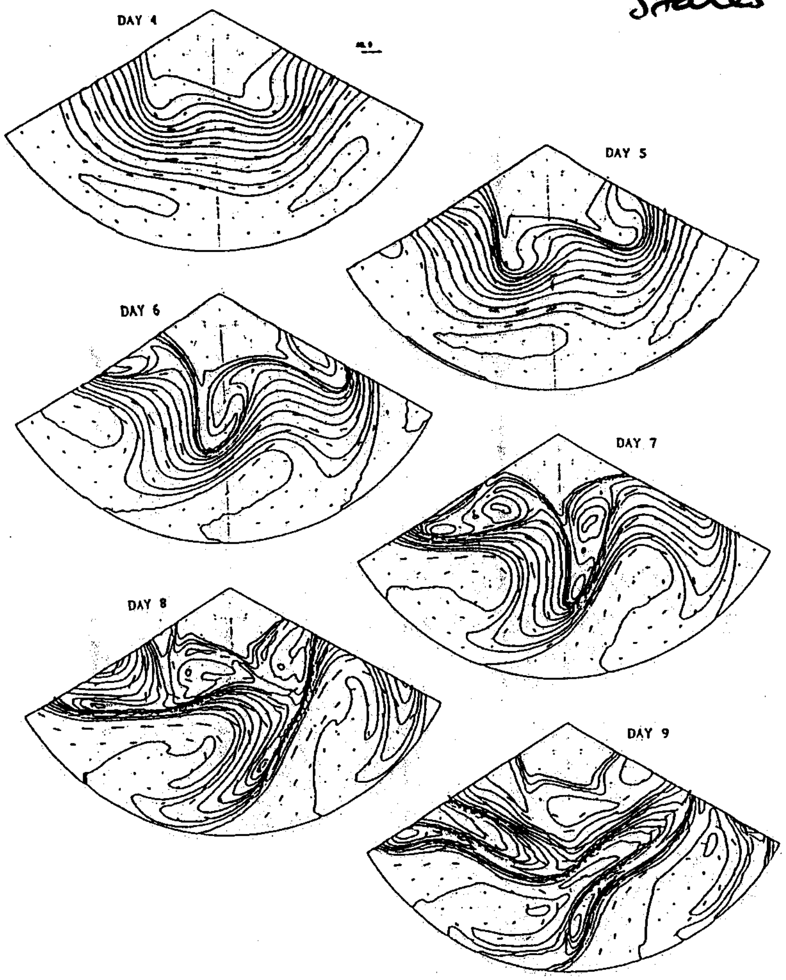
S. (1987 JAS)

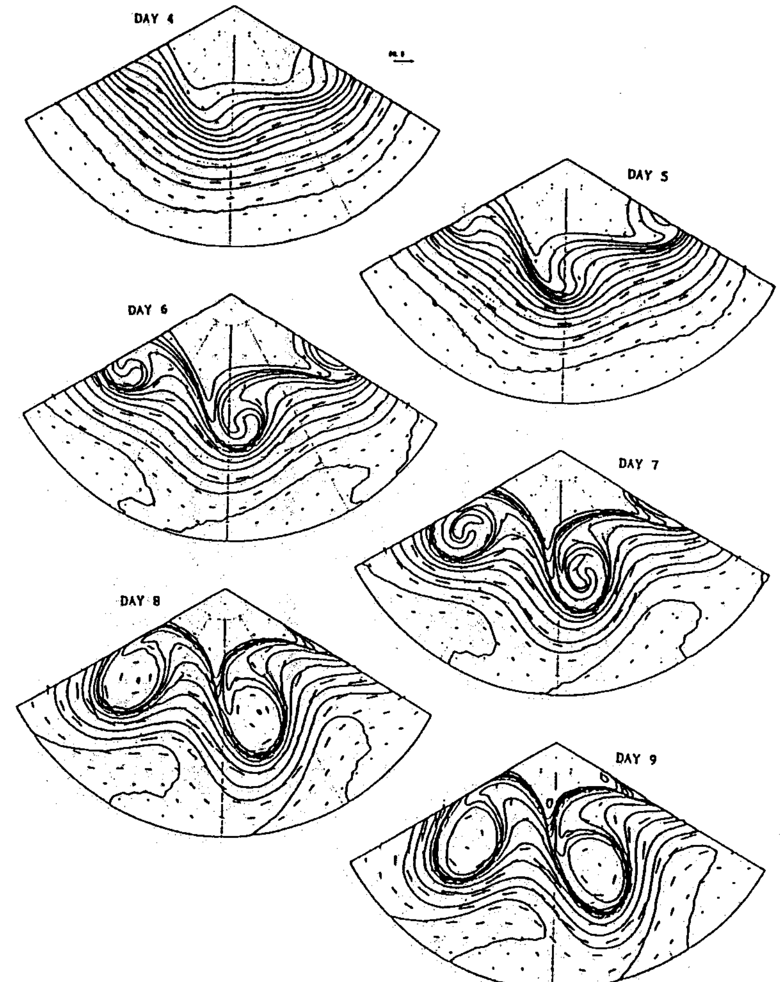
Potential temperature on the PV=2 PVU surface in baroclinic life cycles (wave-6 symmetry imposed)

LC1

The difference is in the basic states (the STJ)

LC2





Thorncroft, Hoskins & McIntyre (1993, QJRMS)

E-P flux divergence in baroclinic life cycles

LC1 LC2

DAY 0

DAY 5

DAY 8

very similar in linear stage of evolution!

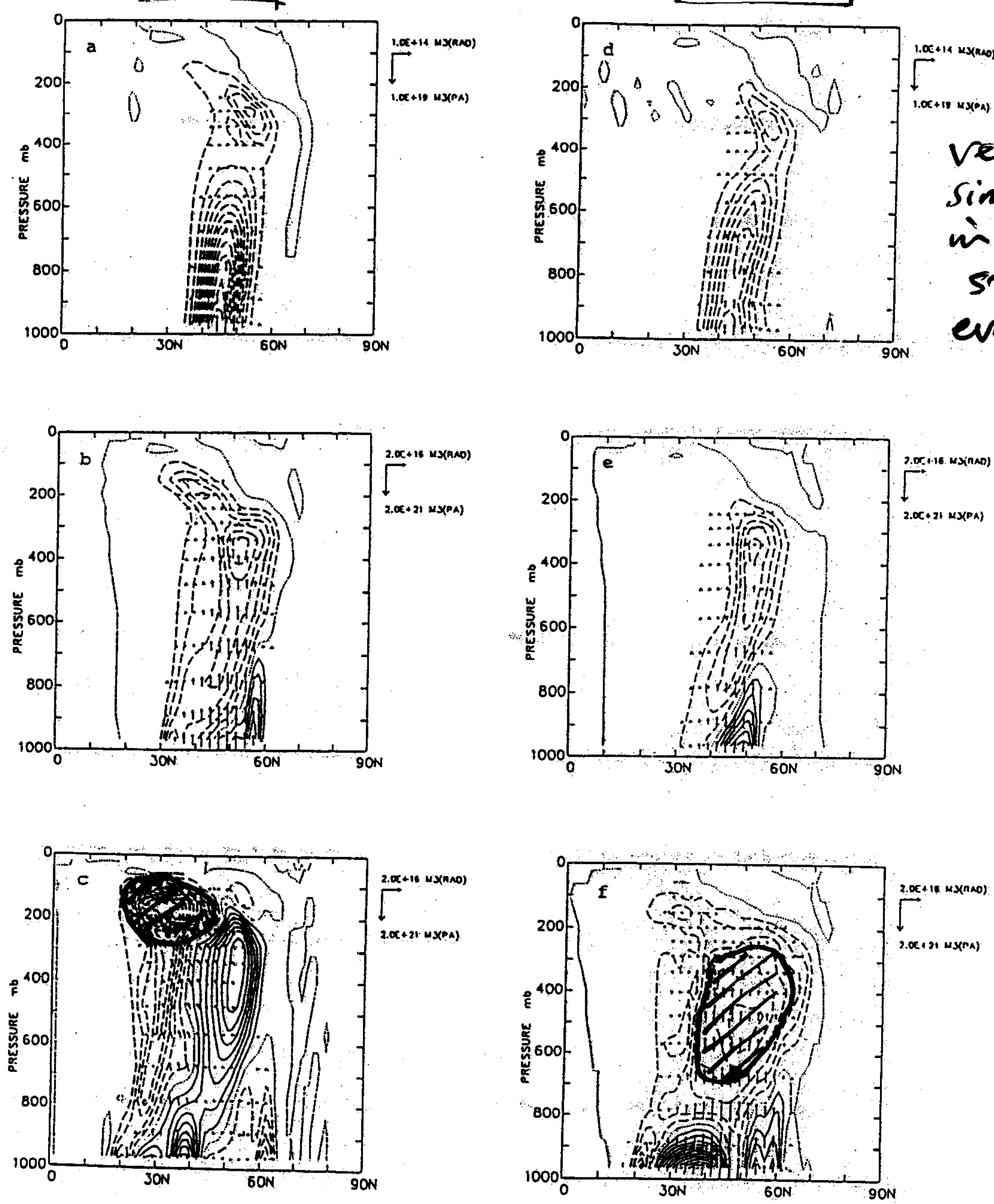


Figure 15. EP flux and its divergence during the two life cycles. (a), (b) and (c) are for days 0, 5 and 8 in LC1 and (d), (e) and (f) are for the same days in LC2. Dashed contours are negative and the contour interval is $4 \times 10^{15}\,m^3$. The divergence in (a) and (d) has been multiplied by 400; note also the different arrow scalings (different by a factor 200), and recall that the surface pressure amplitude is 1 mb at day 0 and that the EP flux and divergence both scale with the amplitude squared. Units are as in Edmon *et al.* It should also be noted that these and subsequent sections overemphasize the vertical contribution to the EP vector because of the stretched ratio of vertical to horizontal scales.

Thorncroft, Hoskins & McIntyre (1993, QJRMS)

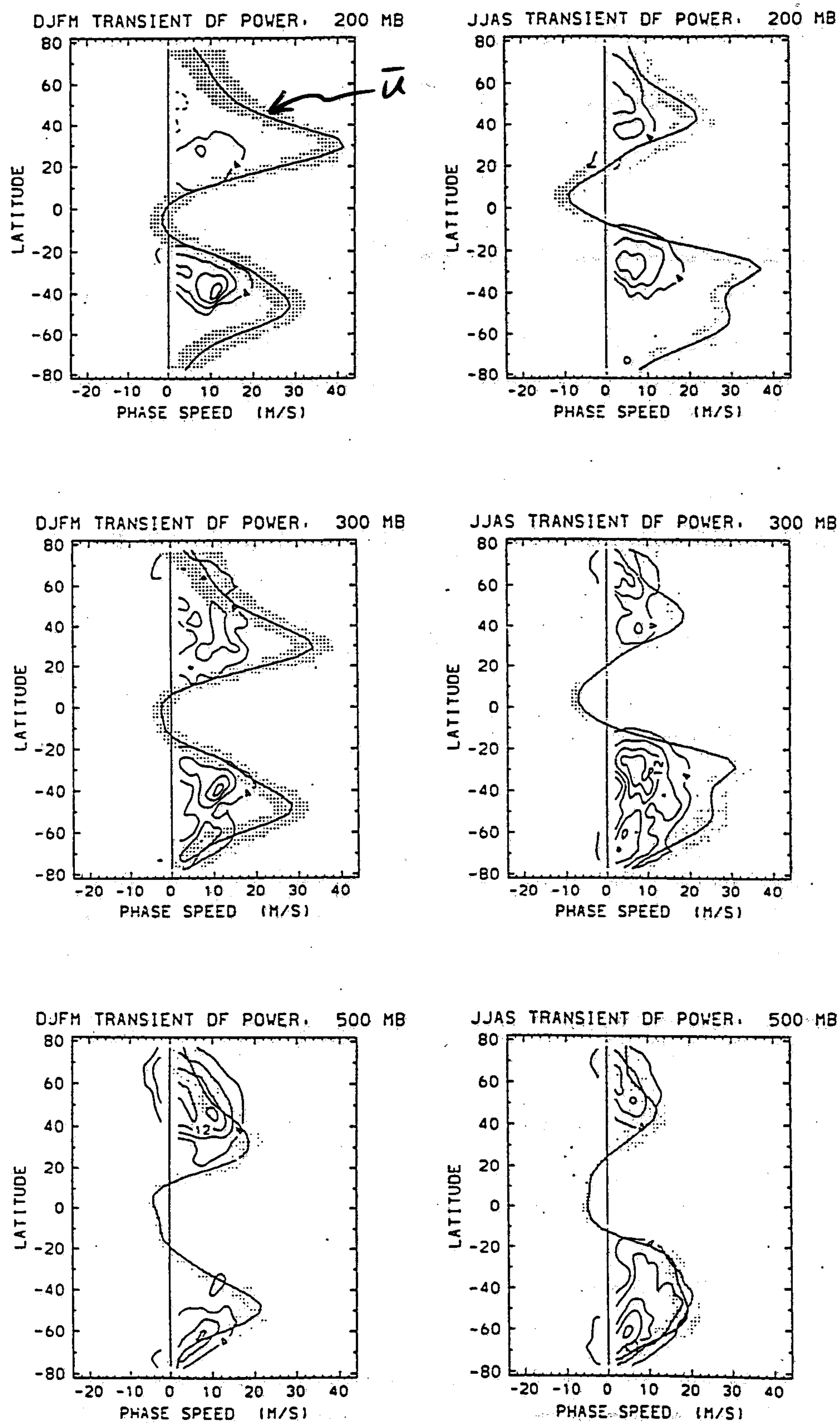
Phase-speed spectra of E-P flux convergence in ECMWF analyses

→ Evidence of LC1 / LC2 ?

200 mb

300 mb

500 mb



LC1 ?

LC2 ?

Randel & Held (1991, JAS)

Phase-speed spectra of $\overline{v'T'}$ and $\overline{u'v'}$ in ECMWF analyses

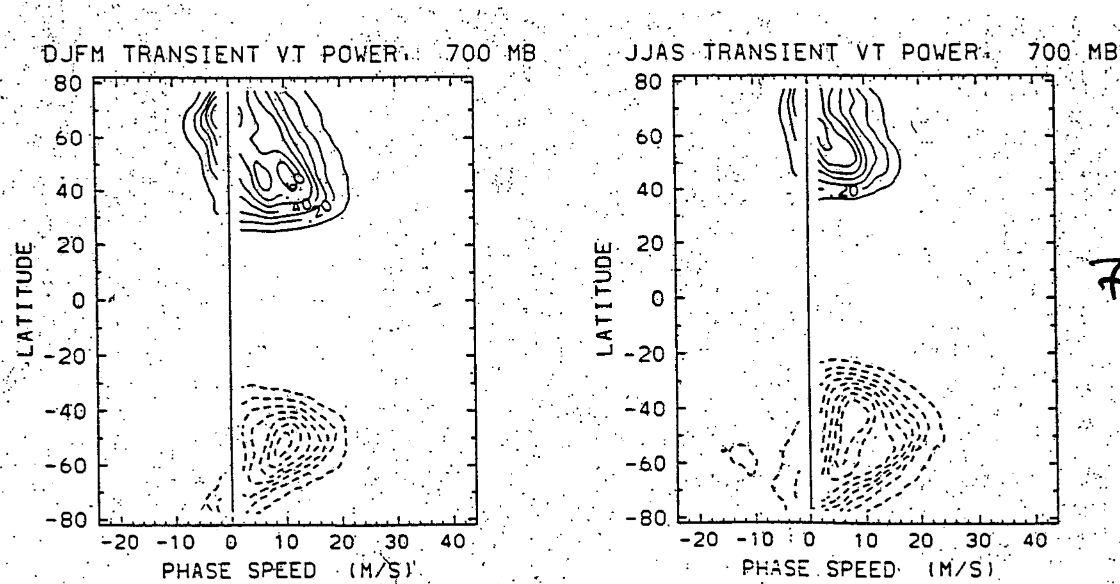


FIG. 5. Contours of 700 mb transient eddy heat flux versus latitude and phase speed for DJFM (left) and JJAS (right). Contour interval is 0.10 K m s^{-1} · Δc^{-1}.

$\overline{v'T'}$ 700 mb

Stronger $\overline{u'v'}$ pattern ⇒ more LC1

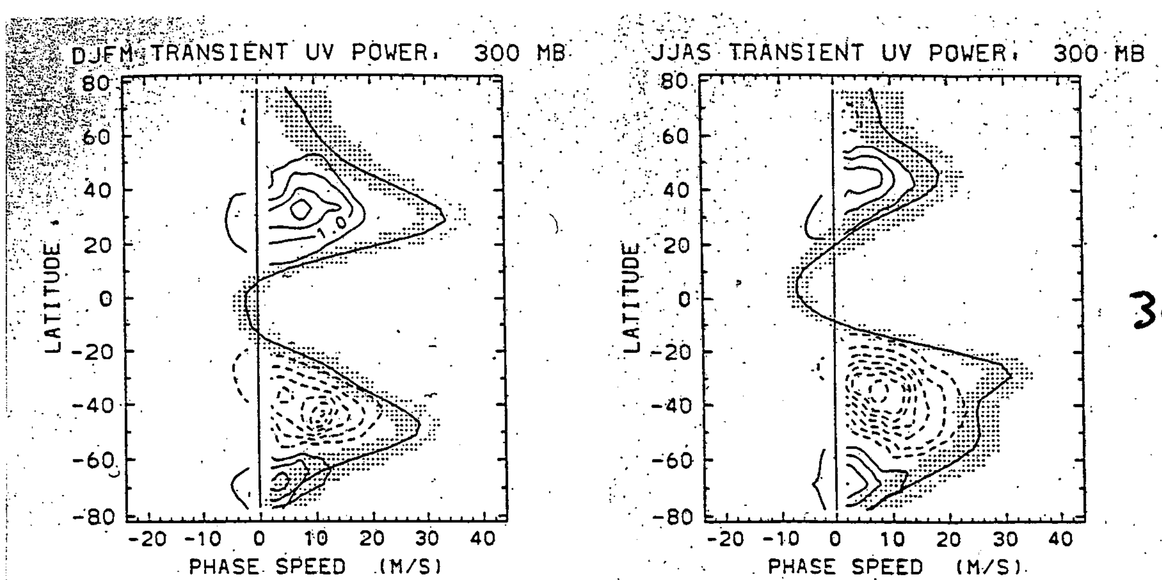


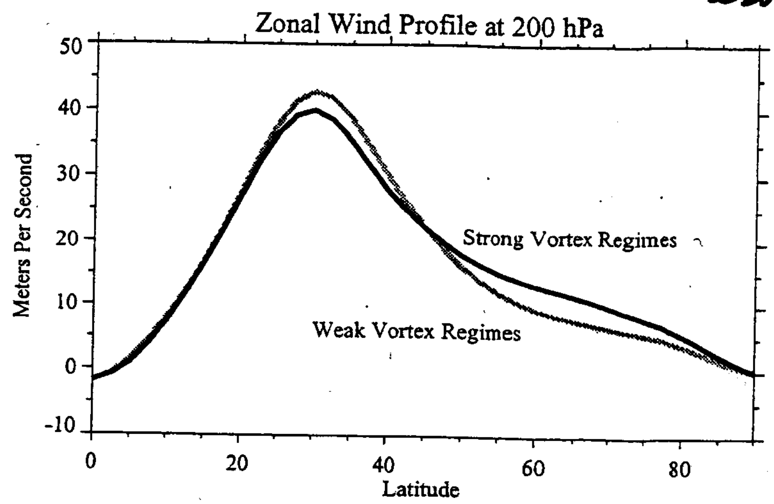
FIG. 6. Contours of 300 mb transient eddy momentum flux versus latitude and phase speed for DJFM (left) and JJAS (right). Contour interval is 0.50 m^2 s^2 · Δc^{-1}, with zero contours omitted. Heavy lines in each panel denote seasonal average zonal mean zonal wind, and shading denotes plus and minus one daily standard deviation.

$\overline{u'v'}$ 300 mb

Randel & Held (1991 JAS)

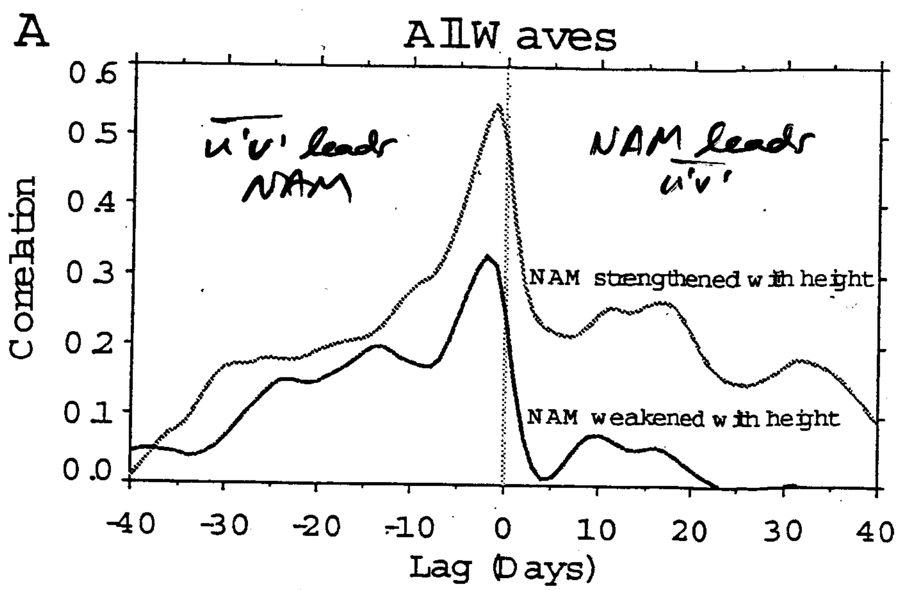
Zonal wind profile at 200 hPa in "strong vortex" and "weak vortex" regimes

⇒ weak vortex regime assoc'd with cyclonic shear



Courtesy of Mark Baldwin (NWRA)

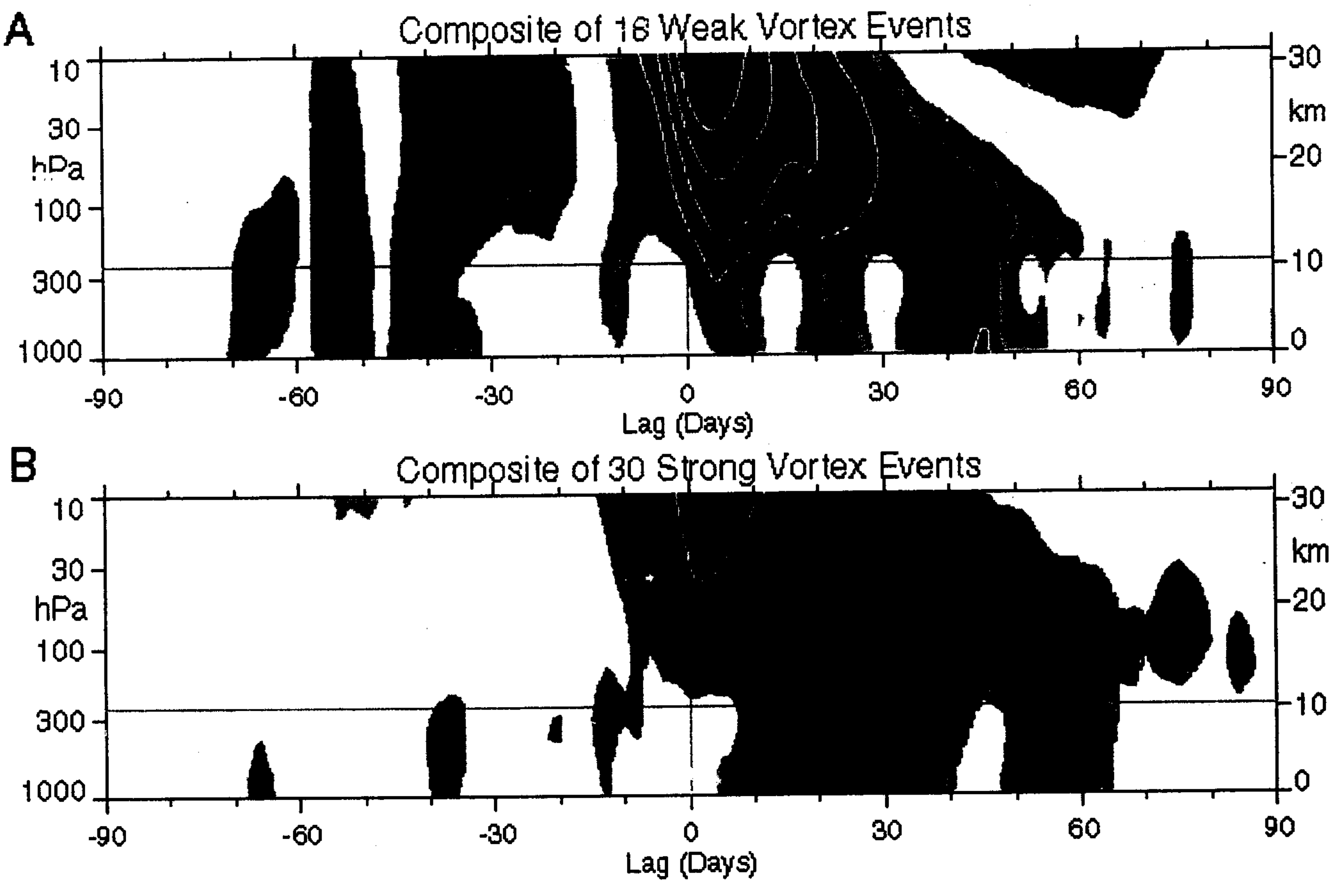
Cross-correlation between 300 hPa NAM and 300 hPa $\overline{u'v'}$ integrated poleward of 20°N



⇒ stratospheric NAM may condition $\overline{u'v'}$ in the upper tropos.

Courtesy of Mark Baldwin (NWRA)

Apparent downward propagation of the Arctic Oscillation index



Baldwin & Dunkerton, Science (2001)