NOTES AND CORRESPONDENCE

Eliassen-Palm Cross Sections for the Northern and Southern Hemispheres

DAVID J. KAROLY

Australian Numerical Meteorology Research Centre, P.O. Box 5089AA, Melbourne, Victoria 3001, Australia

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ABSTRACT

Eliassen-Palm cross sections and residual meridional circulations are presented for the Northern and Southern Hemispheres for summer and winter based on the data of Newell *et al.* (1972, 1974). The cross sections are similar to those presented by Edmon *et al.* (1980) for the Northern Hemisphere except that, in the Southern Hemisphere, there is much less variation between summer and winter than in the Northern Hemisphere.

1. Introduction

Eliassen-Palm (EP) cross sections are meridional cross sections showing the Eliassen-Palm flux **F** by arrows and its divergence by contours. Edmon *et al.* (1980, hereafter EHM) have reviewed the theory and use of EP cross sections as a diagnostic for disturbances on a mean zonal wind. Readers are referred to that paper for all equations, notation and background. They presented EP cross sections based on observational data for the Northern Hemisphere (NH). In this note, EP cross sections are presented based on observations for the Southern Hemisphere (SH).

The best published data set suitable for computing EP cross sections for the SH is the zonal mean basic state and eddy meridional flux data of Newell *et al.* (1972, 1974, hereafter NKVB). This gives data for the zonal-mean total eddy heat flux, rather than the contributions of the stationary and transient eddies separately. Thus, it is possible to compute only EP cross sections based on the total eddy flux data and not for the stationary and transient eddies separately, as in EHM. In fact, the contribution from the transient eddies dominates the total eddy flux statistics for the SH in both summer and winter. The stationary eddy momentum flux is much less than for the transient eddies (Newell *et al.*, 1972) and the stationary eddy heat flux is negligible compared to that for the transient eddies (van Loon, 1980).

The data base for the zonal-mean basic state and total eddy statistics is station data for the period July 1957–June 1963. Time mean and eddy statistics were computed at each station for each month, then these were analyzed on constant pressure surfaces and finally zonal means were calculated. Long-term seasonal means were calculated from the monthly means. Data for the seasons December–February (DJF) and June–August (JJA) are used here. The zonal mean data are given at nine pressure levels (1000, 850, 700, 500, 400, 300, 200, 150 and 100 mb) and at 10° latitude intervals from pole to pole. Data are not available at 1000 mb at 70 and 80°S and at 700, 850 and 1000 mb at 90°S as these positions are below the surface of Antarctica.

To compute the EP cross sections, the quasi-geostrophic expressions for spherical geometry [(3.1) and (3.2) from EHM] are used. All derivatives are calculated using centered finite differences except at the boundaries of the data region, where one-sided finite differences are used. The same graphical conventions as described in EHM are used in the EP cross section figures.

The zonal mean residual meridional streamfunction $\bar{\psi}^*$ is defined by (3.4) and (6.1) in EHM. It is computed by transforming (6.1) to a Poisson equation for $\bar{\psi}^*$, which is solved using a relaxation method. This is a different technique to that described in EHM for computing $\bar{\psi}^*$ and means that the variation of both the meridional and vertical components of the residual circulation determine the streamfunction.

2. Results

The numerical procedures were tested using the Northern Hemisphere data from Oort and Rasmusson (1971, hereafter OR) and the results were compared with those in EHM. At the lower boundary, the differences in the magnitude of $\nabla \cdot \mathbf{F}$ were large at some latitudes, probably due to the use of different finite difference schemes at the boundaries. Else-

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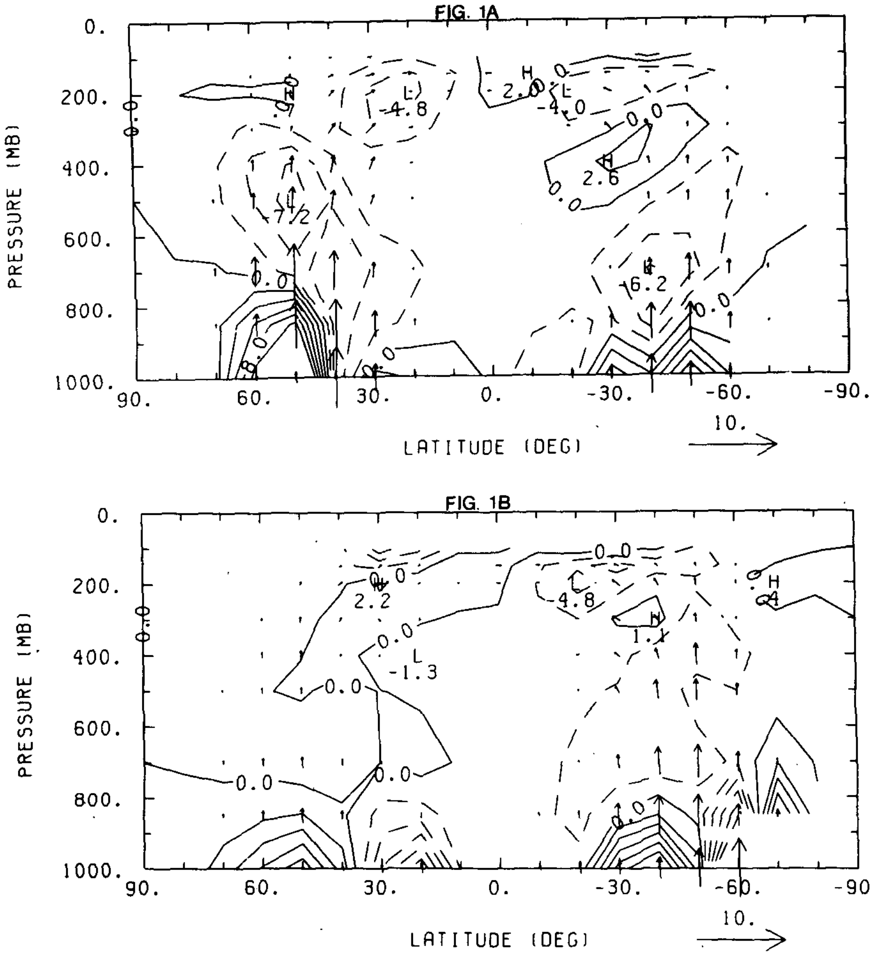

FIG. 1. Eliassen-Palm cross sections for all eddies from the mean data of Newell *et al.* (1972, 1974) for (a) December–February and (b) June–August. The contour interval is 2.0 $\times 10^{15}$ m^3 and negative contours are dashed. The horizontal arrow scale for $\hat{F}_{(\phi)}$ in units of 10^{15} m^3 is indicated at bottom right. A vertical arrow of the same length represents a flux $\hat{F}_{(p)}$, in m^3 kPa, equal to that for the horizontal arrow multiplied by 200 π^{-1} kPa.

where, both the EP cross sections and the meridional streamfunctions were very similar, with the largest differences being less than 10% in magnitude. These differences also can be attributed to the use of different finite difference schemes here and in EHM.

Fig. 1 shows the EP cross sections for all eddies for DJF and JJA calculated using the data from NKVB. The spatial patterns for the NH agree well with the total eddy EP cross sections in EHM calculated from the OR data, except at the lower boundary where the vertical resolution of the NKVB data is coarser than the OR data. In general, the magnitude of $\nabla \cdot \mathbf{F}$ calculated from the NKVB data is smaller than from the OR data as the horizontal resolution of the NKVB data is coarser also. The same scale is used for the EP flux arrows in both seasons and it is apparent that the total eddy fluxes in the NH are much weaker in the summer than in the winter.

In the SH, the spatial pattern of the EP flux is similar to that in the NH but there are some differences. There is much less variation of the magnitudes of $\mathbf{F}$ and $\nabla \cdot \mathbf{F}$ between summer and winter in the SH than in the NH. This reduced seasonal variability in the SH compared with the NH has been noted by Trenberth (1979) in an observational study of the variability of the geopotential height and zonal wind at 500 mb in the SH. In the middle troposphere, $\nabla \cdot \mathbf{F}$ is larger in the SH in summer than winter, which is the opposite seasonal variation to the NH. The magnitude of $\mathbf{F}$ in the SH is slightly larger in JJA than in DJF, consistent with the observed seasonal variation of eddy heat flux in the SH given in van Loon (1980).

Throughout much of the troposphere, there is EP flux convergence, which corresponds with downgradient flux of quasi-geostrophic potential vorticity in regions where the local vorticity gradient is positive.

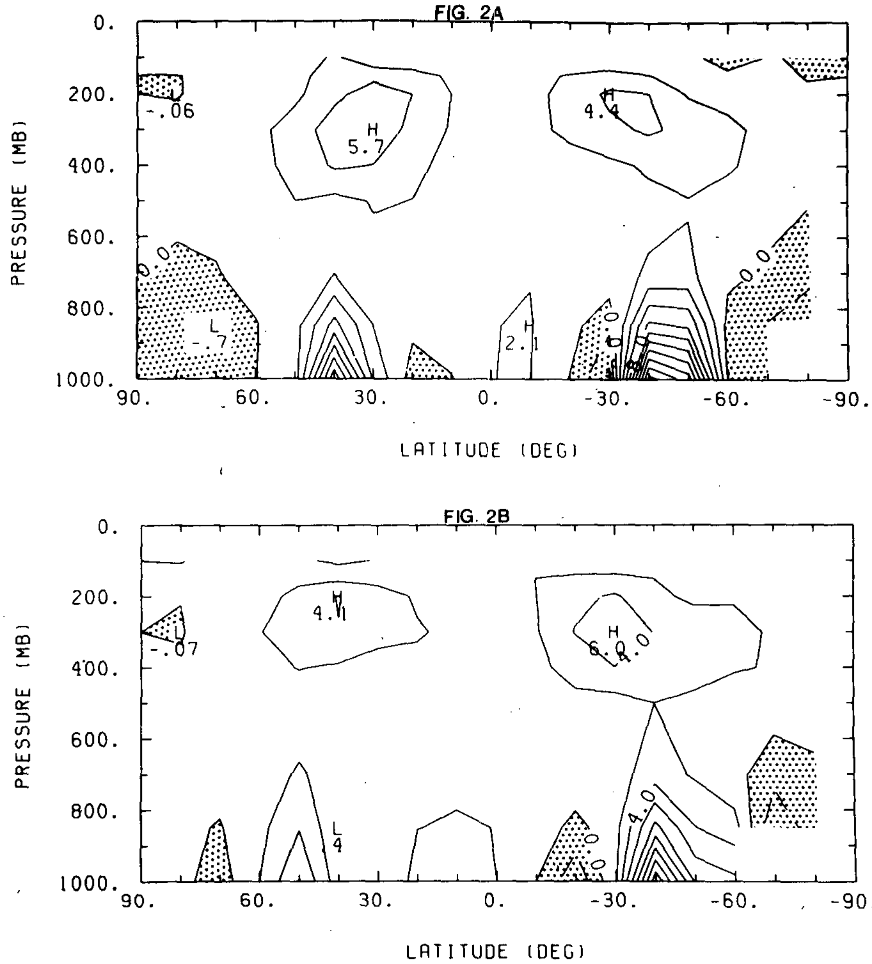

FIG. 2. Nondimensional Mercator quasi-geostrophic potential vorticity gradient β_M for the zonal mean basic state for (a) December–February and (b) June–August. The contour interval is 2.0 and negative values are stippled.

Fig. 2 shows the Mercator vorticity gradient β_M [see Karoly and Hoskins (1982) for definition] for DJF and JJA. β_M is related to the meridional gradient of zonal mean quasi-geostrophic potential vorticity. In general, the vorticity gradient is positive. In fact, the regions of EP flux divergence close to the surface broadly coincide with regions of negative vorticity gradient, so they also correspond to downgradient potential vorticity flux. The difficulties already mentioned with the vertical resolution and the finite difference scheme at the lower boundary for the NKVB data place some doubt on this comparison. A similar comparison has been made between $\nabla \cdot \mathbf{F}$ and β_M for the OR data, where the vertical resolution near the lower boundary is higher, and the agreement of regions of EP flux divergence with negative vorticity gradient is very good. This shows that the quasi-geostrophic potential vorticity flux is generally downgradient in the lower troposphere. This is consistent with the downgradient potential vorticity flux in the nonlinear baroclinic eddy life cycle shown in EHM.

There are regions of EP flux divergence in the upper troposphere and lower stratosphere that are in regions of positive vorticity gradient and show upgradient vorticity flux. Possible causes of this apparent upgradient flux are the nonlinear barotropic decay of baroclinic eddies, diabatic effects associated with deep convection, nonlinear wave interactions, neglected ageostrophic effects, or even that the zonal-mean vorticity gradient may not be representative of the local gradient in regions of eddy activity. Further investigation of these regions of EP flux divergence and the possibility of upgradient flux of quasi-geostrophic potential vorticity is required.

Fig. 3 shows the residual meridional streamfunctions for DJF and JJA. In the NH, the shape of the streamfunction pattern is very similar to that presented in EHM though the magnitudes are slightly

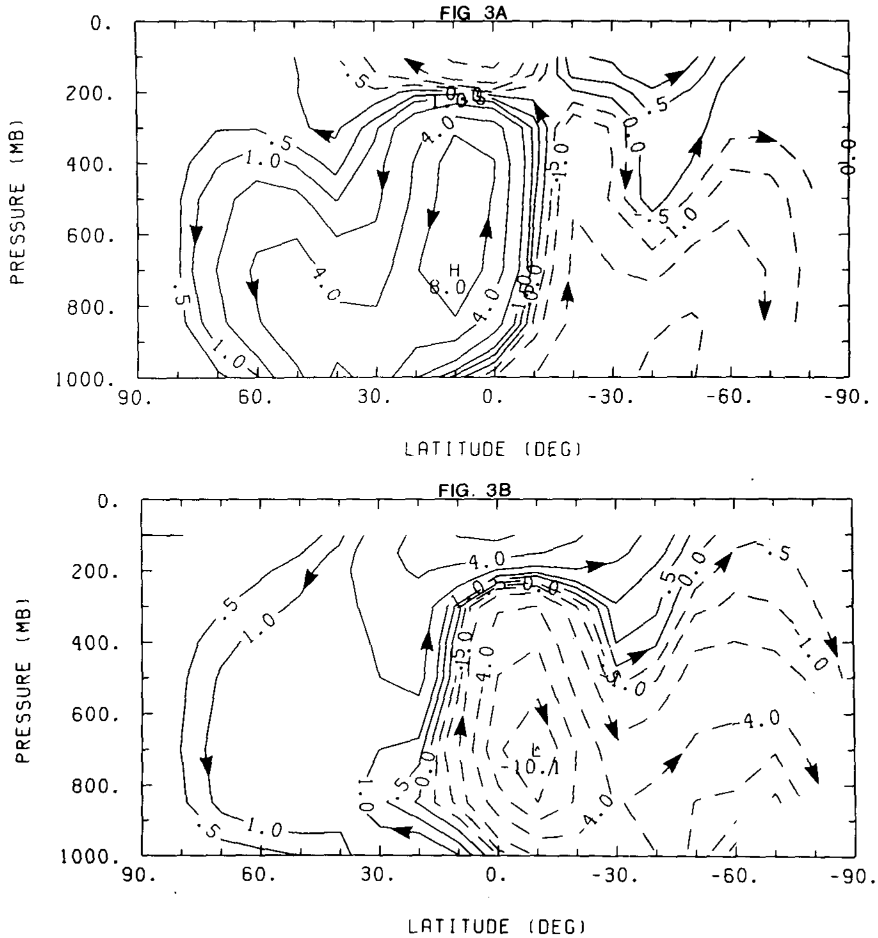

FIG. 3. Residual meridional streamfunction for (a) December–February and (b) June–August. Contours are drawn at values 0, ±0.5, ±1.0, ±2.0, ±4.0, ±6.0, · · · $\times 10^{17}$ m^2 s Pa and arrows indicate the direction of the circulation.

smaller, probably due to the coarser resolution of the NKVB data. As discussed in EHM, this residual meridional circulation is a diabatically forced circulation and is similar to what one would expect for a Lagrangian-mean meridional circulation. In middle latitudes of the winter hemisphere, there is a rising branch of the circulation which is probably due to the ageostrophic effect of the convergence of vertical heat flux of transient eddies. This vertical motion would therefore not be part of a Lagrangian-mean circulation.

In the SH, the residual meridional circulation in the summer is very similar to that in the winter, with a rising branch in middle latitudes. This suggests that the convergence of vertical eddy heat flux is important in the SH summer and that the transient eddies have large amplitude in the SH summer.

From the large-amplitude eddy fluxes and the contribution of the eddies to forcing the meridional circulation, it is apparent that transient eddies play a more important role in the SH summer than they do in the NH summer. The similarities between the EP cross sections and meridional circulations in the SH summer and winter show that there is only small seasonal variation of the eddy amplitude and interaction with the mean flow in the SH, whereas there is large seasonal variation in the NH.

Further computations of EP cross sections and residual meridional circulations for the Southern Hemisphere are in progress using a more recent and comprehensive data set.

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REFERENCES

Edmon, H. J., B. J. Hoskins and M. E. McIntyre, 1980: Eliassen-Palm cross sections for the troposphere. *J. Atmos. Sci.,* **37,** 2600–2616.

Karoly, D. J., and B. J. Hoskins, 1982: Three-dimensional propagation of planetary waves. *J. Meteor. Soc. Japan*, **60** (in press).

Newell, R. E., J. W. Kidson, D. G. Vincent and G. J. Boer, 1972: *The General Circulation of the Tropical Atmosphere and Interactions with Extratropical Latitudes*, Vol. 1. The MIT Press, 258 pp.

——, J. W. Kidson, D. G. Vincent and G. J. Boer, 1974: *The General Circulation of the Tropical Atmosphere and Interactions with Extratropical Latitudes*, Vol. 2. The MIT Press, 371 pp.

Oort, A. H., and E. M. Rasmusson, 1971: *Atmospheric Circulation Statistics*. NOAA Prof. Pap. No. 5, U.S. Dept. of Commerce, Washington, DC, 323 pp.

Trenberth, K. E., 1979: Interannual variability of the 500 mb zonal mean flow in the Southern Hemisphere. *Mon. Wea. Rev.*, **107**, 1515–1524.

van Loon, H., 1980: Transfer of sensible heat by transient eddies in the atmosphere on the Southern Hemisphere: An appraisal of the data before and during FGGE. *Mon. Wea. Rev.*, **108**, 1774–1781.

Curvature Diminution in Equatorial Wave, Mean-Flow Interaction[1]

TIMOTHY J. DUNKERTON[2]

Department of Atmospheric Sciences, University of Washington, Seattle 98195

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ABSTRACT

It is shown that slowly varying linear equatorial Rossby-gravity waves in a barotropically neutral mean-wind profile near the equator accelerate the mean flow in a stabilizing sense there. This indicates that the Rossby-gravity wave, believed to be the driving force in the easterly acceleration phase of the quasi-biennial oscillation, cannot force a barotropically unstable mean flow near the equator. Mean flows generated near the equator in the easterly phase of the oscillation in the context of these approximations will therefore resemble or be approximately bounded by a parabola of curvature β, where β is the planetary vorticity gradient. This result does not depend upon a "barotropic adjustment" process, although the latter has been suggested in the past and would yield the same result, but over a broader latitudinal area.

1. Introduction

The theory of equatorial waves and wave, mean-flow interaction has an interesting history in recent years. Probably the most notable achievement of the theory to date is a successful qualitative explanation of the quasi-biennial oscillation advanced by Holton and Lindzen (1972). According to these authors, the stratospheric mean-wind oscillation arises from the alternating absorption, *via* radiative damping, of the lowest order equatorial wave modes—the Kelvin and Rossby-gravity waves—which are believed to be excited in the troposphere. The Kelvin wave appears to drive the mean flow in a westerly direction when the Rossby-gravity wave is shielded by the low-level flow, while in the opposite phase the Rossby-gravity wave drives the mean flow in an easterly sense similarly. Holton and Lindzens' theory explains this phenomenon in terms of linear, slowly varying waves while accounting for their influence on the mean flow in a "quasi-linear" model. Despite these limitations the theory has been greatly supported by the laboratory simulation of Plumb and McEwan (1978), who were also able to explain their laboratory oscillation in terms of a similar theory.

The laboratory simulation involved a stratified, nonrotating fluid in a cylindrical annulus having internal waves forced from below in the form of a standing wave. This standing wave can be regarded as the sum of two internal waves of equal amplitude propagating in opposite directions. Plumb and McEwan demonstrated that this configuration is unstable with respect to a quasi-biennial-type oscillation. In the laboratory, viscous diffusion provided the necessary absorption mechanism. The resulting descending mean-shear zones formed a remarkable parallel to the observed oscillation.

However, there is at least one important difference between the observed and laboratory oscillations, and it is the purpose of this note to explore this difference theoretically. Because the atmosphere is in rotation, there is an equatorial waveguide formed by the Coriolis force near the equator. Unlike the laboratory simulation this leads to a significant meridional

[1] Contribution No. 622, Department of Atmospheric Sciences, University of Washington.

[2] Present affiliation: National Center for Atmospheric Research, PO Box 3000, Boulder, CO 80307.

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