The Effect of Transient Eddies on the Stationary Eddy Isobaric Height Field

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ABSTRACT

The effect of transient eddies (TE) on stationary eddies (SE) is studied in terms of how the TEs force the SE isobaric height variance in the northern extratropics. The method used correlates the observed SE height field with the TE geopotential tendencies, which describe the net forcing effect that arises from the TE fluxes of heat and momentum and from the associated secondary circulations. Diagnostic estimates are made based on atmospheric circulation statistics for February 1979 and for an ensemble of several winters. The results for the TE forcing of the SE isobaric height variance are compared with those of the SE potential enstrophy.

Although the net forcing effect of all TEs on the SE potential enstrophy is a damping one at all levels, the TEs tend to maintain the SE isobaric height variance in the lower troposphere. This effect is caused by the high-frequency synoptic-scale TEs. The low-frequency TEs, by contrast, tend to damp the SE isobaric height variance at all levels. The sensitivity of the method to uncertainties in the input data is estimated by making calculations from different datasets valid for the same period.

The methods applied in this paper might be used as a diagnostic tool in the validation of general circulation models.

1. Introduction

A main characteristic of the atmospheric circulation is its large variability in time. This "large-scale turbulence," which is vividly illustrated in any sample of consecutive weather maps, has an important effect on the time-mean circulation.

An arbitrary quantity can be represented as the sum of contributions from the zonally averaged time-mean flow, stationary eddies (SE) and transient eddies (TE). In this paper we consider the dynamic effect of the TEs on the SEs; this effect will be referred to as (TE, SE) interaction.

The intensity and characteristics of both the TEs and the SEs depend upon the sampling duration. If, for example, several winters are considered, the SE part of the flow describes the longitudinal variation of the climatological flow and has a smaller magnitude than the TE part, which arises from TEs of many frequencies. If, on the other hand, the sampling period is short, the SE part also contains the large-scale, slowly varying features of the flow and typically has the same magnitude as the TE part.

Many diagnostic studies on the (TE, SE) interaction show that the TEs tend to damp the SE total energy (available potential energy + kinetic energy) integrated over the whole mass of the atmosphere (Holopainen 1970; Lau and Oort 1982). Holopainen et al. (1982) showed that, in terms of quasi-geostrophic potential enstrophy also, the TEs exert a damping effect on the SEs at all pressure levels. This effect was shown to arise mainly from the low-frequency transient eddies. Besides total energy and potential enstrophy, the temperature variance at the lower boundary is a dynamically important quantity (e.g., Hoskins et al. 1985). Some diagnostic studies (e.g., Lau 1979; Holopainen 1983a) indicate that the TEs also dampen the SE temperature variance at the lower boundary during winter in the Northern Hemisphere.

Despite the results mentioned above, it can be misleading to describe the climatic (TE, SE) interaction solely as a damping process. Holopainen and Oort (1981) showed that in the extratropics the TEs maintain the barotropic part of the wintertime SEs. They also showed that the TEs are a mechanism counterbalancing the effect of surface stress in the vertically integrated budget of atmospheric vorticity in the regions of the centers of action, such as the North Pacific low and the North Atlantic low. Because the mean surface stress is related to the mean geostrophic wind at the surface, the TEs are thus important for determining the distribution of the time-mean pressure and circulation in the extratropical lower troposphere. Describing the (TE, SE) interaction only as damping process would therefore be misleading.

Youngblut and Sasamori (1980) studied the time-mean potential vorticity budget in the midtroposphere over the Northern Hemisphere in January 1963. Their Figs. 2b and 4b indicate damping of the SE potential

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enstrophy in this case too, where the sampling duration is only 1 month. Holopainen and Fortelius (1987a) used four different approaches to describe the net effect of high-frequency TEs on a 10-day mean flow. It was concluded that the high-frequency TEs are important in forcing lower-frequency fluctuations. These high-frequency eddies appeared to be particularly important in the distribution of the (10-day) mean circulation in the lower troposphere in the vicinity of the storm tracks.

It is obvious from these studies that the appearance of the (TE, SE) interaction depends on the physical quantity and time scale of the transient eddies considered.

In this paper we study the (TE, SE) interaction in terms of how the TEs affect the SE isobaric height variance over the Northern Hemisphere. We find this approach more enlightening than the one based on the SE potential enstrophy, for which some parallel results are presented. The principal period studied is February 1979, the main full month of Special Observing Period I of the Global Weather Experiment (or FGGE). We have made calculations using several datasets, all of which should, in principle, yield the same results. Differences in the results obtained from different datasets give some information about the sensitivity of the applied method to uncertainties in data. Besides the February 1979 case, the (TE, SE) interaction in the case of the wintertime climatological flow is discussed.

Section 2 examines the theoretical framework used. The datasets used in the diagnostic part of this study are described in section 3. The results for the SE potential enstrophy budget are presented in section 4; the SE isobaric height variance in section 5; and a physical interpretation of the results is given in section 6. In section 7, the results are summarized and the potential applicability of the methods (e.g., in validation of atmospheric general circulation models) is discussed.

2. Theoretical considerations

The net forcing effect of eddies on a mean flow consists of the sum of the direct effects in the form of the convergence of eddy fluxes of heat and momentum, and of the indirect effects produced by the eddy-induced secondary circulations.

One method of studying the net TE forcing effect on the time-mean flow is to calculate the tendencies which a quasi-geostrophic atmosphere would experience at the moment when the TE fluxes of heat and momentum were "switched on." This approach was introduced by Lau and Holopainen (1984). It was recently applied by Mullen (1987) and Holopainen and Fortelius (1987a) as a method of illustrating the effect of high-frequency TEs in blocking flows.

In this section we first present the basic equations of the tendency method. We then derive from these equations the expressions used in diagnosing the TE, SE interaction.

a. *The tendency method*

The dynamics of the time-mean motion is, in the quasi-geostrophic framework, determined by the time-averaged potential vorticity equation

$$\partial\bar{q}/\partial t = -\bar{\mathbf{V}}\cdot\nabla\bar{q} + (\partial\bar{q}/\partial t)_H + (\partial\bar{q}/\partial t)_{\mathrm{TE}} + (\partial\bar{q}/\partial t)_F \quad (1)$$

where, as usual, q denotes quasi-geostrophic potential vorticity, $\mathbf{V}$ geostrophic wind and the bar denotes a time average. The right-hand side (rhs) terms with subindices H, TE and F refer to the forcing due to heating, TEs and friction. The TE forcing is given by

$$(\partial\bar{q}/\partial t)_{\mathrm{TE}} = -\nabla\cdot\overline{q'\mathbf{V}'} \quad (2)$$

where the prime denotes a deviation from the time-mean.

The vertical boundary conditions are provided by the thermodynamic energy equation, which at the lower boundary reads

$$\partial\bar{\theta}/\partial t = -\bar{\mathbf{V}}\cdot\nabla\bar{\theta} + \bar{H} + (\partial\bar{\theta}/\partial t)_{\mathrm{TE}} + S\bar{\omega}_M + S\bar{\omega}_F. \quad (3)$$

Here θ denotes potential temperature and H the effect of heating. The static stability, S, is a function of pressure only and $\omega = dp/dt$. The terms ω_M and ω_F are the lower boundary vertical velocities due to mountains and Ekman layer friction, respectively; at the upper boundary these terms do not exist. The TE forcing term in (3) is given by

$$(\partial\bar{\theta}/\partial t)_{\mathrm{TE}} = -\nabla\cdot\overline{\theta'\mathbf{V}'}. \quad (4)$$

If the distribution of $\nabla\cdot\overline{q'\mathbf{V}'}$ in the interior of the atmosphere and $\nabla\cdot\overline{\theta'\mathbf{V}'}$ at the boundaries is known, the three-dimensional field of the associated geopotential tendency $(\partial\bar{\Phi}/\partial t)_{\mathrm{TE}}$ can be calculated by inversion (see Lau and Holopainen 1984 and Holopainen and Fortelius 1987a for the details of the method.) The same holds true for the other terms in (1) and (3). Symbolically, the inverted form of (1) and (3) can be written as an equation for the geopotential tendency

$$\partial\bar{\Phi}/\partial t = (\partial\bar{\Phi}/\partial t)_{\mathrm{ADV}} + (\partial\bar{\Phi}/\partial t)_H + (\partial\bar{\Phi}/\partial t)_{\mathrm{TE}} + (\partial\bar{\Phi}/\partial t)_M + (\partial\bar{\Phi}/\partial t)_F \quad (5)$$

where Φ [$=gZ$] is the geopotential (g is the acceleration of gravity and Z the isobaric height). The left-hand side (lhs) term is the actual geopotential tendency, which is zero in steady state conditions. The term $(\partial\bar{\Phi}/\partial t)_{\mathrm{ADV}}$ arises from the mean-flow advection terms in (1) and (3). (In the classical zonally averaged case considered by Pfeffer 1981, 1987 this advection term is absent). The forcing term due to mountains $(\partial\bar{\Phi}/\partial t)_M$ entirely due to boundary forcing.

As discussed by Lau and Holopainen (1984) and by Holopainen and Fortelius (1987a), $(\partial\bar{\Phi}/\partial t)_{\mathrm{TE}}$ appears to be a convenient way of illustrating the net TE forcing effects. The effect on temperature and geostrophic wind can be obtained from $(\partial\bar{\Phi}/\partial t)_{\mathrm{TE}}$ by using the hydrostatic

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and geostrophic relationships. A zonal average of the geostrophic wind forcing would give the two-dimensional (axisymmetric) zonal wind tendencies discussed by Pfeffer (1981, 1987).

b. The net TE forcing effect on the SE potential enstrophy and isobaric height variance

If one multiplies (1) by $\bar{q}^*$ and takes an area average over the isobaric surface one gets

$$\partial/\partial t\{\tfrac{1}{2}\bar{q}^{*2}\} = \{\bar{q}^*(\partial\bar{q}^*/\partial t)_{\mathrm{TE}}\} + (\cdots) \qquad (6)$$

where { } denotes an area average on an isobaric surface and ( )* a deviation from the zonal average. Only the term arising from the TEs is written on the rhs of (6); similar terms naturally would also arise from the other terms on the rhs of (1).

The SE potential enstrophy, $\frac{1}{2}\{\bar{q}^{*2}\}$, is a quadratic measure of the intensity of the SEs. The term $\{\bar{q}^*(\partial\bar{q}^*/\partial t)_{\mathrm{TE}}\}$ on the rhs of (6) is the rate of change of the SE potential enstrophy due to the TE forcing effect. This forcing term can be written in the form

$$\{\bar{q}^*(\partial\bar{q}^*/\partial t)_{\mathrm{TE}}\} = \{(\overline{q'\mathbf{V}'})_D \cdot \nabla\bar{q}^*\} \qquad (7)$$

where $(\overline{q'\mathbf{V}'})_D$ is the divergent part of $\overline{q'\mathbf{V}'}$. [For details see Holopainen and Fortelius (1987b) who discussed the conversion of potential enstrophy from the time-mean flow to the TEs: $\{\bar{q}^*(\partial\bar{q}^*/\partial t)_{\mathrm{TE}}\}$ is a part of that conversion].

A normalized measure of the TE forcing effect on the SEs is given by the ratio

$$\nu^q = -\{\bar{q}^*(\partial\bar{q}^*/\partial t)_{\mathrm{TE}}\}/\{\tfrac{1}{2}\bar{q}^{*2}\}. \qquad (8)$$

This ratio can be called the SE potential enstrophy damping coefficient due to the (TE, SE) interaction. The inverse of the coefficient gives the time scale for the interaction. If the coefficient is negative, the TEs tend to maintain the SE potential enstrophy.

If one multiplies (5) by $g^{-2}\bar{\Phi}^*$ and takes an area average one gets

$$\partial/\partial t\{\tfrac{1}{2}\bar{Z}^{*2}\} = \{\bar{Z}^*(\partial\bar{Z}^*/\partial t)_{\mathrm{TE}}\} + (\cdots). \qquad (9)$$

The isobaric height variance in SEs, $\frac{1}{2}\{\bar{Z}^{*2}\}$, is another quadratic measure of the intensity of the SEs. It is perhaps a more easily understandable quantity than $\frac{1}{2}\{\bar{q}^{*2}\}$. No simple relationship exists between these quantities because the height field is influenced not only by potential vorticity but also by boundary temperature.

Equation (7) gives an explicit expression for $\{\bar{q}^*(\partial\bar{q}^*/\partial t)_{\mathrm{TE}}\}$ in terms of divergent TE potential vorticity fluxes and SE potential vorticity gradients. Expressions analogous to (6)–(7) can be written for the SE temperature variance at the lower boundary pressure surface. Yet no simple explicit expression can be written for $\{\bar{Z}^*(\partial\bar{Z}^*/\partial t)_{\mathrm{TE}}\}$.

In analogy with (8) we define the isobaric height variance damping coefficient

$$\nu^Z = -\{\bar{Z}^*(\partial\bar{Z}^*/\partial t)_{\mathrm{TE}}\}/\{\tfrac{1}{2}\bar{Z}^{*2}\} \qquad (10)$$

where $(\partial\bar{Z}^*/\partial t)_{\mathrm{TE}}$, and thus ν^Z, may be written as a sum of two parts, one associated with TE heat fluxes and the other with the TE vorticity fluxes.

Expressions (6)–(10) form the basic framework for the present analysis.

3. Datasets and procedures used

The TE forcing of the SE potential enstrophy and SE isobaric height variance have been calculated using atmospheric circulation statistics from various periods. One of these periods is February 1979, the main full month of Special Observing Period I of the Global Weather Experiment (FGGE). Most calculations are based on the first (often referred to as "main") version of the FGGE (Level III-b) analyses of meteorological variables, prepared by ECMWF (European Center for Medium Range Weather Forecasts), GLA (Goddard Laboratory for Atmospheres) and GFDL (Geophysical Fluid Dynamics Laboratory). However, the latest version, referred to as "FINAL", of the global, initialized, FGGE level III-b analyses from ECMWF was also used.

The "MAIN" dataset from ECMWF was further used to create a filtered time series for February 1979 (strictly speaking, for the period 3–26 February). The filter applied is a high-pass (HP) filter which isolates TEs with periods of less than about 6 days. It originates from Lau and Lau (1984). Using the filtered time series, ν^Z_{HP} was computed both for the entire month and for the 10-day period of 16–25 February. The latter period was characterized by enhanced amplitudes of the SEs, and by enhanced HP-eddy induced tendencies in the vicinity of the principal storm tracks.

For the case of wintertime climatological flow two datasets were used. The first one (hereafter referred to as OORT data) covers the winter seasons of the 5-year period 1968–73. It is part of the larger data library described by Oort (1983). It is based on the statistics of observations at the rawinsonde stations around the world. The second dataset (hereafter referred to as NMC data, e.g., Lau et al. 1981) is based on twice-daily analyses of the atmospheric circulation to the north of 20°N during the 11 winters 1965/66–1975/76, and prepared by the U.S. National Meteorological Center.

Besides the unfiltered time series, the NMC dataset contains covariance statistics of the time series obtained by applying the following filters (Blackmon 1976):

- a low-pass (LP) filter retaining fluctuations between 10 days and a season
- a band-pass (BP) filter retaining fluctuations with periods between 2.5 and 6 days.

The LP time series appear to describe (Blackmon 1976) mainly phenomena of a large horizontal scale while the BP time series are related to smaller-scale (referred

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to in the following as synoptic-scale or cyclone-scale) phenomena.

Expression (7) was used to estimate $[\bar{q}^*(\partial\bar{q}^*/\partial t)_{\mathrm{TE}}]$. The TE potential vorticity flux divergences and the associated divergent fluxes were estimated from the horizontal TE fluxes of heat and momentum as described in appendix B of Holopainen and Fortelius (1987b). The method described in the Appendix of Holopainen and Fortelius (1987a) was used to calculate the forcing term $(\partial\bar{\Phi}/\partial t)_{\mathrm{TE}}$.

4. The TE effect on the SE potential enstrophy in February 1979

The framework for the analysis of the general circulation of the atmosphere in terms of potential enstrophy was discussed by Holopainen and Fortelius (1987b), who studied the maintenance of the TE potential enstrophy $(\frac{1}{2}\{q'^2\})$ over the Northern Hemisphere in February 1979. The following discussion on the SE potential enstrophy $(\frac{1}{2}\{\bar{q}^{*2}\})$ complements the analysis reported in that paper. This section forms a background for the main new results to be presented in section 5.

The vertical profile of $\frac{1}{2}\{\bar{q}^{*2}\}$ (Fig. 1a) has a maximum at around 300–400 hPa. The values of $\{\bar{q}^*(\partial\bar{q}^*/\partial t)_{\mathrm{TE}}\}$ (Fig. 1b) are seen to be negative at all levels implying conversion of potential enstrophy from the SEs to the TEs. Formally, the TEs thus damp the SE potential enstrophy, especially in the upper troposphere. The conversion of potential enstrophy from the SEs to the TEs appears to represent about half of the total conversion of potential enstrophy from the time-mean flow to the TEs in February 1979 (Holopainen and Fortelius 1987b); the other half is converted from the zonally averaged time-mean flow. The damping effect of the TEs on the SE potential enstrophy seen in Fig. 1b is in qualitative agreement with the results of Youngblut and Sasamori (1980) for January 1963, and with those from the winter climate SE studies by Holopainen et al. (1982) and Holopainen (1983a,b).

Figure 1c shows the vertical profile of the damping coefficient ν^q, obtained from the profiles in Figs. 1a and 1b. The time scale of the damping, given by the inverse of the coefficient, is shortest at 700 hPa, where it is about 9 days. This is much larger than the value (4–5 days) found when studying the wintertime climate (Holopainen et al. 1982; Holopainen 1983a,b). A natural explanation for this is that the low-frequency TEs present in the case of many winters are only partly present in the data for one single month.

A qualitative interpretation of the positive ν^q is obtained by noting that the air over the extratropical continents in winter has higher time-mean static stability and potential vorticity than the air over the warm oceans. In the low-frequency TEs, the air has enough

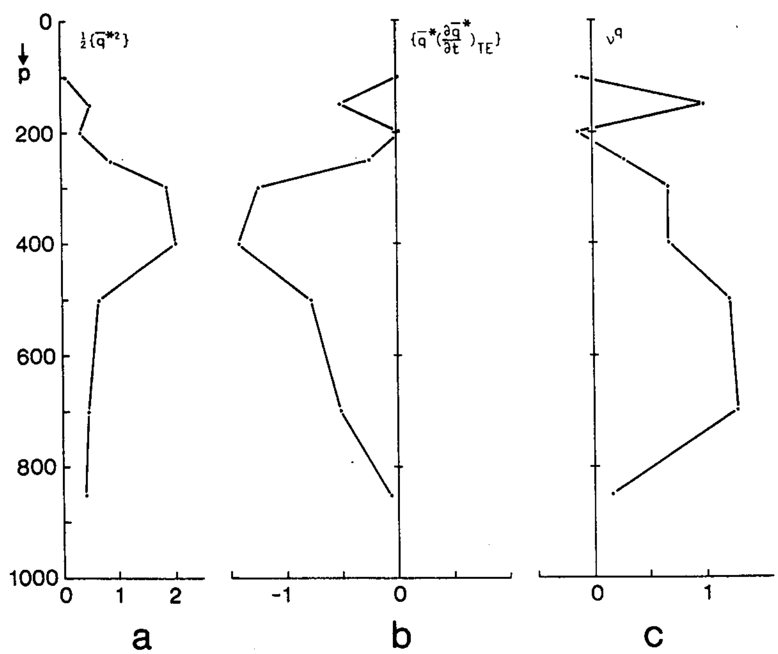

FIG. 1. The vertical distribution of (a) the stationary eddy (SE) potential enstrophy $\{\frac{1}{2}(q^{*2})\}$ (unit: 10^{-9} s^{-2}); (b) the net transient eddy (TE) effect on the SE potential enstrophy $\{[\bar{q}^*(\partial\bar{q}^*/\partial t)_{\mathrm{TE}}]\}$ (unit: 10^{-15} s^{-3}); (c) $\nu^q = -\{\bar{q}^*(\partial\bar{q}^*/\partial t)_{\mathrm{TE}}\}/\{\frac{1}{2}\bar{q}^{*2}\}$ (unit: 10^{-6} s^{-1}). (Positive values indicate damping of the SE potential enstrophy by the TEs). The area-average is taken over the northern extratropics (20°–90°N). The estimates are based on the "MAIN" (FGGE Level III-b) datasets prepared by ECMWF for February 1979.

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time to adjust its properties to the local conditions dictated by the underlying surface and radiation conditions. Thus, these eddies can smoothen the spatial differences of the potential vorticity and produce downgradient (west-east) potential vorticity flux.

5. The TE effect on the isobaric height variance of the stationary eddies

In this section we consider the TE effect on the SE isobaric height variance in the framework of (9)–(10). In subsection 5*a* we first compare, for February 1979, the results for the SE isobaric height variance with the corresponding results presented in section 4 for the SE potential enstrophy. In addition to the effect of all TEs, the contribution by high-pass (HP), synoptic-scale eddies is analyzed separately.

In order to get some ideas about the generality of the results for one single month we consider in subsection 5*b* the (TE, SE) interaction in the case of the wintertime climate. In addition to the effect of all TEs, we present results for the contributions by low-pass (LP) and band-pass (BP) eddies.

This section presents the formal results and discusses their reliability; the major part of the physical interpretation of the results will be given in section 6.

a. February 1979

Figure 2 shows the vertical profiles of $\frac{1}{2}\{\bar{Z}^{*2}\}$, $\{\bar{Z}^{*}(\partial\bar{Z}^{*}/\partial t)_{\rm TE}\}$ and ν^{Z} for the northern extratropics (20°–90°N) in February 1979 from the MAIN and FINAL datasets prepared by ECMWF. Figure 2b shows that, in this framework, the effect of the TEs is a maintaining one in the lower troposphere and a damping one in the upper troposphere. The coefficient ν^{Z} (Fig. 2c) differs radically from ν^{q} (Fig. 1c). At 850 hPa, a forcing time scale of 3–4 days is implied while the damping in the upper troposphere has a time scale of about two weeks.

The results obtained from the MAIN and FINAL datasets prepared by ECMWF (Fig. 2) are seen to be almost identical. Figure 3 shows the forcing coefficient ν^{Z} calculated for February 1979 using the "MAIN" datasets from ECMWF, GLA and GFDL. In principle, all the curves should be identical. The basic features of the TE forcing, i.e., strengthening of the SEs in the lower troposphere and weakening in the upper troposphere, are indeed seen to be alike. Considerable quantitative differences are found, however. The results from ECMWF and GLA data are, in general, rather similar to each other but deviate from those based on the GFDL data.

Figure 4 shows the coefficient $\nu^{Z}_{\rm HP}$ obtained by calculating $(\partial\bar{Z}^{*}/\partial t)_{\rm TE}$ in (10) from the high-pass (HP) data described in section 3. The solid line refers to the whole month of February 1979. The separate effects of the heat and vorticity fluxes (not shown) have the same sign in the lower troposphere, where they tend to force the SEs, and oppose each other in the upper troposphere, where, however, the maintaining effect of

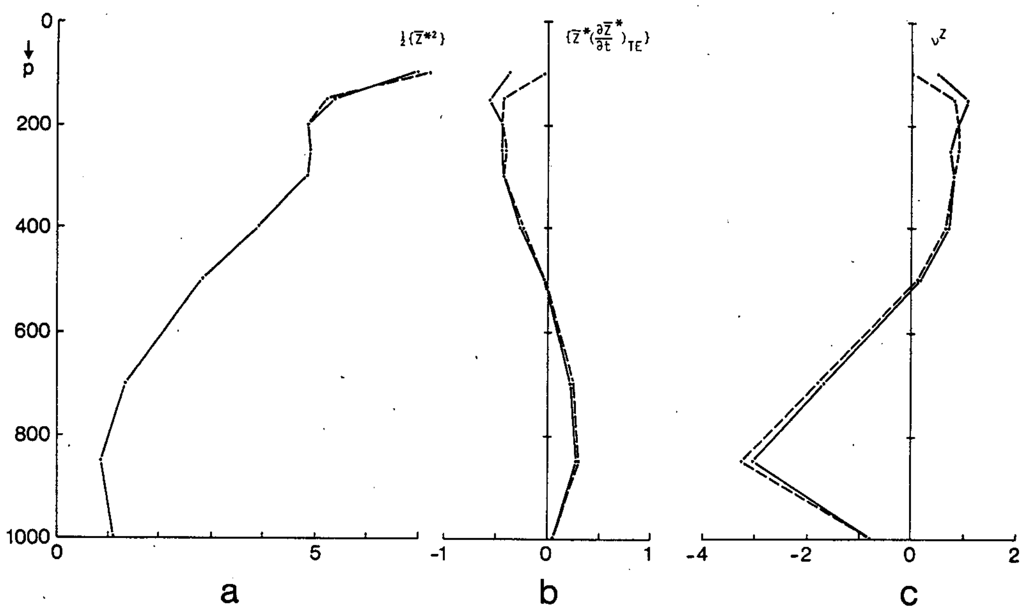

FIG. 2. The vertical distribution of (a) the (half) SE isobaric height variance $[\frac{1}{2}(\bar{Z}^{*2})]$ (unit: 10^{3} m^{2}); (b) the net TE effect on the SE isobaric height variance $\{\bar{Z}^{*}(\partial\bar{Z}^{*}/\partial t)_{\rm TE}\}$ (unit: 10^{-3} m^{2} s^{-1}); (c) $\nu^{Z} = -\{\bar{Z}^{*}(\partial\bar{Z}^{*}/\partial t)_{\rm TE}\}/\{\frac{1}{2}\bar{Z}^{*2}\}$ (unit: 10^{-6} s^{-1}). (Positive values indicate damping). The area-average is taken over the northern extratropics (20°–90°N). Solid and dashed lines refer to the "MAIN" and "FINAL" (FGGE Level III-b) datasets, respectively, prepared by ECMWF for February 1979.

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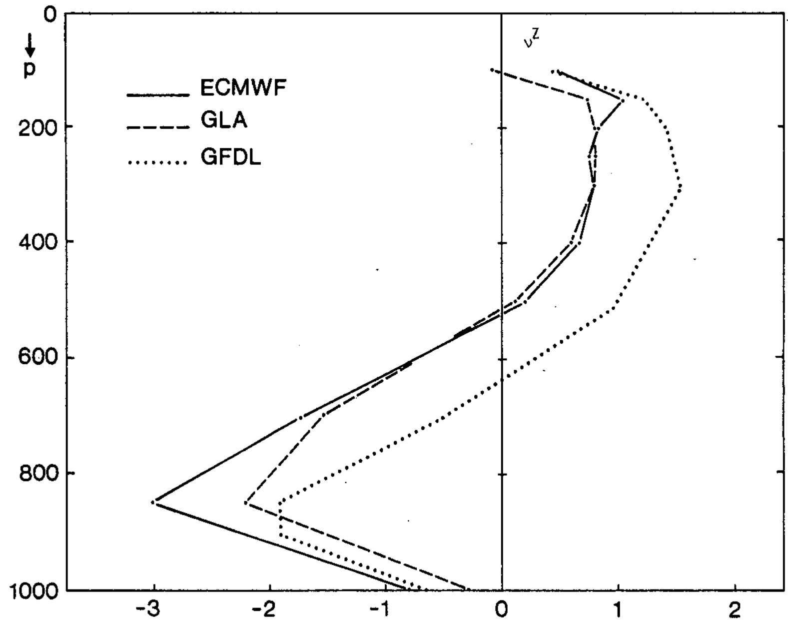

FIG. 3. The coefficient ν^Z, as in Fig. 2c, but estimated from the "MAIN" datasets prepared by ECMWF (solid line), GLA (dashed line) and GFDL (dotted line) for February 1979. Unit: $10^{-6}\ s^{-1}$.

the vorticity flux dominates. While the coefficient for all TEs (Fig. 2c) changes sign in the vertical, the coefficient for the HP eddies exhibits positive forcing at all levels, with a maximum at 850 hPa.

The dashed line in Fig. 4 refers to the 10-day period of 16–25 February 1979. During this "blocking" period, the absolute value of the damping coefficient ν^Z_{HP} is much larger than during the rest of the month.

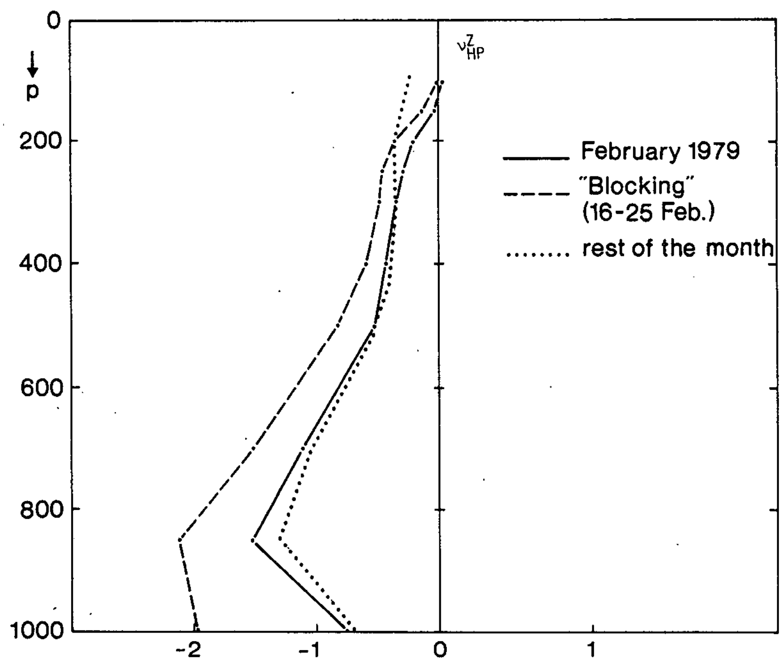

FIG. 4. The coefficient $\nu^Z_{HP} = -\{\bar{Z}^*(\partial\bar{Z}^*/\partial t)_{HP}\}/\{\frac{1}{2}\bar{Z}^{*2}\}$, calculated from the high-pass (HP) filtered data for February 1979 (solid line). The dashed line refers to the period 16–25 February 1979 characterized by a blocking-type flow and the dotted line to rest of the month.

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This is mainly due to differences in the vorticity flux contribution between the two periods. The greater magnitude of $\nu_{\rm HP}^{Z}$ is compatible with the results of Hansen and Sutera (1984), who found an enhanced nonlinear cascade of kinetic energy from cyclone-scale waves to longer waves during blocking months.

On the basis of Figs. 2–4, we may attribute the negative values of ν^Z in Figs. 2c and 3 to the effect of cyclone-scale eddies. The positive values in the upper troposphere must be related to other mechanisms, presumably lower-frequency TEs.

b. Wintertime climate

Results for the SE isobaric height variance in winter are shown in Fig. 5. Both the OORT and NMC datasets indicate a strong damping effect by the TEs in the upper troposphere (Fig. 5b), and a maintaining effect in the lower troposphere. In this respect the results for February 1979 are qualitatively similar to those for the winter climate.

The coefficient ν^Z (Fig. 5c) in the upper troposphere is about twice as large as in February 1979 and implies a damping time scale of about 6 days. In the lower troposphere large quantitative differences are seen between the results derived from the OORT and NMC datasets, the time scales of the forcing e.g., at 850 hPa being 3 and 7 days, respectively. A closer analysis reveals that the differences arise mainly from differences in the TE vorticity fluxes. The differences between the basic circulation statistics in these two datasets have been discussed by Lau and Oort (1981, 1982).

Figure 6 shows the coefficients $\nu_{\rm LP}^{Z}$ and $\nu_{\rm BP}^{Z}$ obtained by calculating $(\partial\bar{Z}^*/\partial t)_{\rm TE}$ in (10) from low-pass and band-pass NMC datasets, respectively. The two curves are seen to be radically different from each other. The net effect of the low-frequency TEs on the SE isobaric height variance is a damping one at all levels, while that of the synoptic-scale TEs is a maintaining one. Coefficient $\nu_{\rm BP}^{Z}$ is seen to have its largest magnitude in the lower troposphere. By comparing $\nu_{\rm LP}^{Z}$ and $\nu_{\rm BP}^{Z}$ with ν^Z (Fig. 5c), one notices that, obviously, the BP eddies determine the sign of the effect of all TEs in the lower troposphere while the LP eddies dominate in the upper troposphere.

It is interesting to note that the effect of the BP eddies in the climatological case (Fig. 6) is practically the same as that of the HP eddies in February 1979 (Fig. 4). It appears that both the HP and the BP datasets essentially describe cyclone-scale transients. The effect of these transients seems to be rather a robust feature of the general circulation.

As in the case of Fig. 4, the values of ν^Z in Fig. 6 can be written as a sum of two contributions, one from the eddy vorticity fluxes, the other from the eddy heat fluxes. In the BP eddies, these two effects (not shown) have the same sign in the lower troposphere, and counterbalance each other to some extent in the upper troposphere (as one would expect on the basis of Fig. 3 in Lau and Holopainen 1984). The tendency associated with the heat flux appears to dominate in the lower troposphere and that with the vorticity flux above the 850 hPa level, as in the case of the HP eddies in February 1979.

For the LP eddies the contribution of the vorticity fluxes to the tendency fields is much larger than that

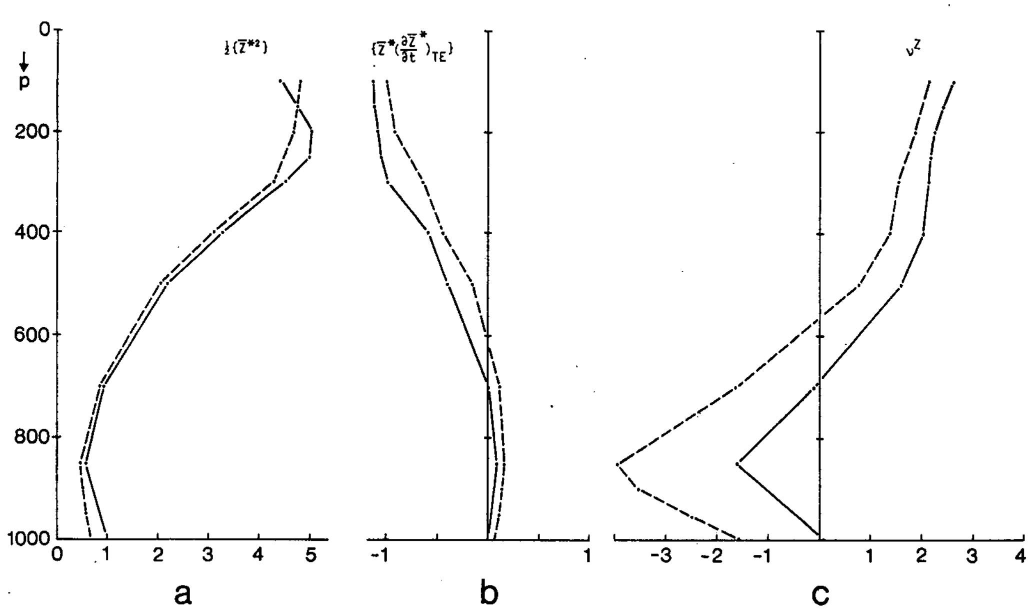

FIG. 5. As in Fig. 2 but for an ensemble of several winters obtained from the NMC (solid lines) and OORT (dashed lines) datasets.

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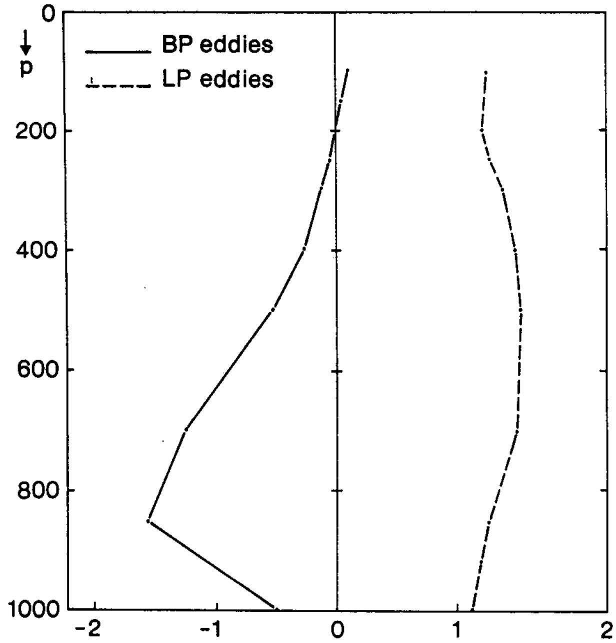

FIG. 6. The coefficient ν^Z, as in Fig. 5c, but obtained from the low-pass (LP) and band-pass (BP) filtered NMC datasets.

due to the heat fluxes, and this contribution is almost independent of height. (See Fig. 5 in Lau and Holopainen 1984.) The profile of ν^Z_{LP} in Fig. 6 of the present study appeared indeed to be determined by the vorticity fluxes in the LP eddies, except at the 1000 hPa level.

The results presented above refer to average isobaric conditions in the whole extratropics (20°–90°N). The contributions of different latitudinal zones (not shown) to these averages indicate that the results presented in Figs. 2–6 are reliable, in the sense that they are not small residuals of large positive and negative values arising from different zones.

6. Interpretation

Figure 7 shows the geopotential height and tendency fields at two pressure levels (300 and 850 hPa). One can see that at both levels there is a negative correlation between the $\bar{Z}^*$ and $(\partial\bar{Z}^*/\partial t)_{LP}$ fields, while the correlation between $\bar{Z}^*$ and $(\partial\bar{Z}^*/\partial t)_{BP}$ patterns is positive. This is in accord with the results shown in Fig. 6.

The sign of the LP eddy forcing at 300 hPa is practically everywhere opposite to that of $\bar{Z}^*$. The damping effect is seen to be especially strong at high latitudes. The pattern of $(\partial\bar{Z}^*/\partial t)_{BP}$ has a larger amplitude at 850 hPa (panel 7f) than at 300 hPa (panel 7e). As discussed by Lau and Holopainen (1984), in the lower troposphere, the net effect of cyclone-scale eddies is a tendency to generate low pressure on the cyclonic side and high pressure on the anticyclonic side of a storm track. This effect is seen at 850 hPa as the dipole structures that tend to maintain the standing waves in connection with the two main storm tracks over the western part of the oceans. The large positive values of $(\partial\bar{Z}^*/\partial t)_{BP}$ in panel 7f in the Eurasian sector around latitude 50°N arise from the asymmetric distribution of the two principal storm tracks along this latitude.

The area-averaged isobaric height variance is roughly proportional to the area-averaged kinetic energy. The results for the SE isobaric height variance must then be compatible with those for the SE kinetic energy. Both theoretical and observational studies (Simmons et al. 1983; Holopainen 1984; Wallace and Lau 1985) have shown that the energy of the low-frequency transients over the Northern Hemisphere in winter at least partly originates from the climatological SEs by means of the barotropic instability of the wavy time-mean flow. The positive values of the damping coefficient ν^Z_{LP} in Fig. 6 are, presumably, a reflection of this instability. According to observations, the cyclone-scale TEs feed kinetic energy not only to the zonal mean flow but also to the SEs (Wallace and Lau 1985). The negative values of ν^Z_{BP} in Fig. 6 (and of ν^Z_{HP} in Fig. 4) are compatible with this.

Negative viscosity phenomena have been discussed extensively in connection with the mean meridional momentum fluxes (e.g., Starr 1968). A nonlinear cascade of kinetic energy from cyclone-scale eddies to longer waves is one aspect of these phenomena. Momentum fluxes have their largest magnitude in the upper troposphere and lower stratosphere while, according to this study, the negative damping is especially a feature of the lower troposphere. The reason for this apparent discrepancy is the Coriolis force associated with the eddy-induced ageostrophic circulation which is the main contributor to the net eddy force in the lower troposphere.

A comparison of Figs. 1c and 2c shows that the TE effect on potential enstrophy is diffusive ($\nu^q > 0$) and that the effect on isobaric height variance and, hence, on kinetic energy is not diffusive ($\nu^Z < 0$ in the lower troposphere). This difference is mainly related to the stability term in the expression for potential vorticity. It was first shown by Wiin-Nielsen and Sela (1971) that the eddy flux of potential vorticity associated with this term is directed towards smaller values of the mean potential vorticity. Another reason for the difference between ν^q and ν^Z is that the latter incorporates the TE effect on the boundary temperature, whereas the former does not.

7. Summary and discussion

In the wintertime Northern Hemisphere, the TEs tend to increase the isobaric height variance of the stationary eddies in the lower troposphere and to reduce it in the upper troposphere. There appears to be a large difference between the effect of the TEs of different time scales. The higher-frequency, synoptic-scale TEs tend to increase the SE isobaric height variance at all

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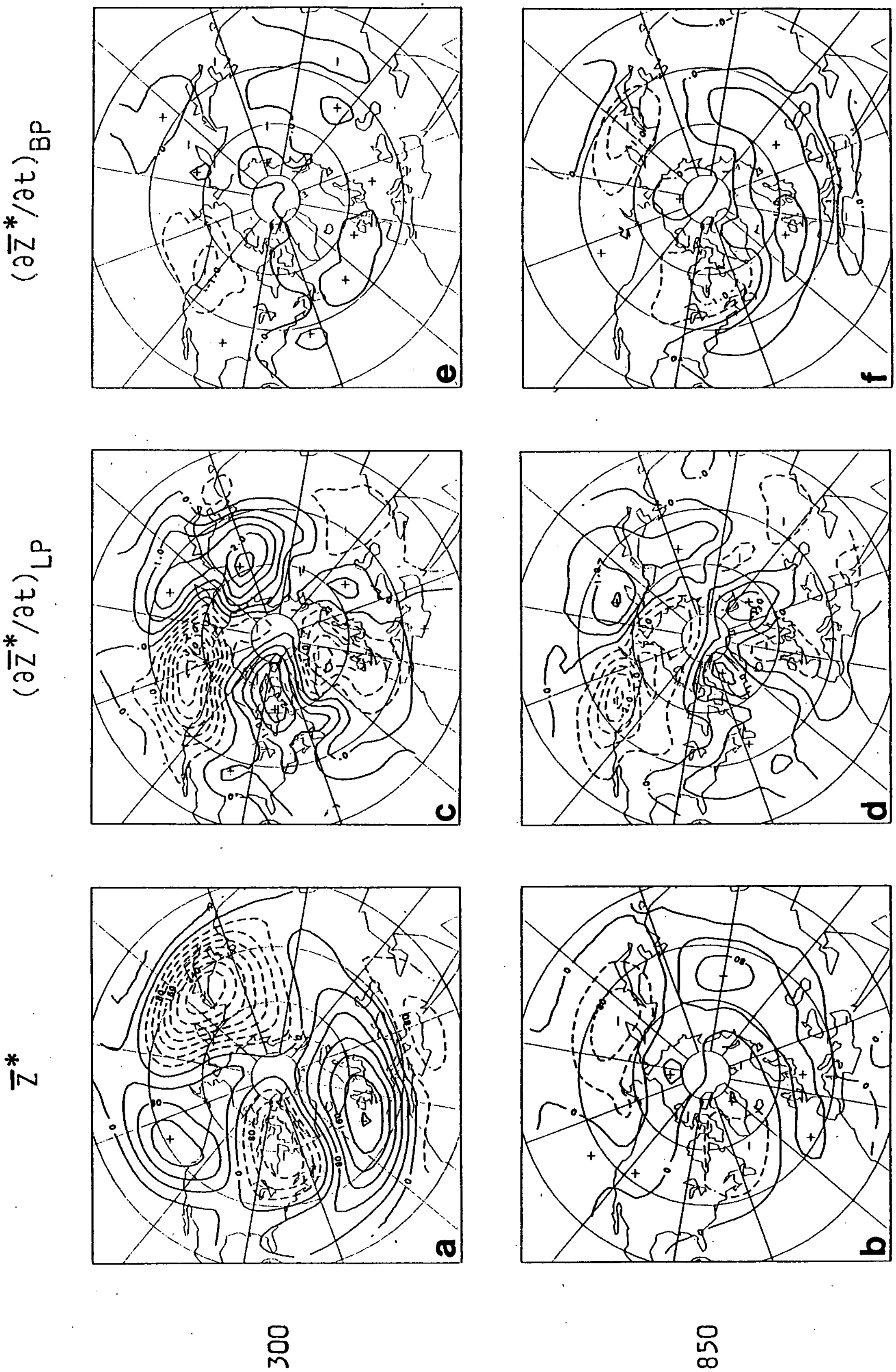

FIG. 7. The $\bar{Z}^*$ field (left), $(\partial\bar{Z}^*/\partial t)_{LP}$ field (center) and $(\partial\bar{Z}^*/\partial t)_{BP}$ field (right) for 300 hPa (upper panels) and 850 hPa (lower panels). The fields are based on the NMC dataset for an ensemble of 11 winters. Isoline spacing is 40 m for the height fields and 0.5×10^{-4} m s^{-1} for the tendency fields.

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levels and determine the net effect of all TEs in the lower troposphere. The low-frequency, large-scale TEs damp the SE isobaric height variance at all levels and determine the net effect in the upper troposphere.

This method might be useful in the validation of general circulation models. The vertical profile of ν^Z, determined from observations and also from model output could, for example, provide a check on the ability of a particular model to simulate the interaction between the SEs and TEs. In low-resolution models the high-frequency, cyclone-scale eddies cannot be explicitly resolved. In light of our results, an adequate simulation of the time-mean flow in the lower troposphere with such models still requires that the forcing effect of the cyclone-scale eddies is properly parameterized.

Because of the geostrophic wind relationship, a study of the isobaric height variance gives more weight to higher latitudes than a study of kinetic energy. This could be avoided in future applications of the present approach by considering the isobaric streamfunction instead of the isobaric height.

Because no explicit expression can be written for the geopotential tendency, the profile of ν^Z can be used only as a diagnostic tool, not in the basic formulation of SE models. For the latter purpose ν^q, and the corresponding coefficient of the SE temperature damping at the lower boundary, must be used.

The definition of the SEs includes a zonal average. Therefore, even if one can formally write, e.g., ν^Z as a sum of local contributions, the physical interpretation of such contributions is not clear. Moreover, the pattern of the product of $\bar{Z}^*$ and $(\partial\bar{Z}^*/\partial t)_{\rm TE}$ changes sign when either one does. One way to elucidate the local forcing effect of high-frequency eddies on the lower-frequency components of the atmospheric circulation would be to study the relationship between the isobaric height anomalies (for example, the deviation of the monthly-mean isobaric height field from the long-term average for that month) and the corresponding anomalies of the tendencies due to the BP (or HP) eddies. Such a study is, however, beyond the scope of the present paper.

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