Transient Eddy Forcing of the Rotational Flow during Northern Winter

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ABSTRACT

The total direct transient eddy forcing of the vorticity balance in the upper troposphere during northern winter is examined using 11 years of 2 to 8-day bandpassed global analyses. Most examinations of the importance of high-frequency eddy vorticity fluxes to the maintenance of either the climatological standing waves or low-frequency anomalous flows have focused on only the horizontal flow or the rotational component of the horizontal transient flow. The latter simplification has been shown to be questionable for planetary scales. The vorticity flux by the transient divergent flow produces a forcing of the mean streamfunction that is of comparable magnitude with the equivalent rotational term. However, the streamfunction forcing by the sum of the transient vertical advection and twisting terms largely balances the forcing by the vorticity flux convergence associated with the divergent flow. The result is that the convergence of the eddy vorticity flux by the total transient flow is not a good approximation to the total forcing of the long-term mean streamfunction by the high-frequency eddies. These results are quantified for the mean northern winter season November–March 1980/81–1990/91.

The respective roles of each transient eddy term in the vorticity equation in maintaining two large-scale, low-frequency anomalous flows are also examined. One case involves a pronounced circulation anomaly that persisted for more than a decade over the North Pacific, and the second case relates to the maintenance of extremes of the North Atlantic Oscillation. In both cases, transient vorticity fluxes systematically reinforce and help to maintain the upper-tropospheric streamfunction anomalies. Just as for the climatological standing waves, a consideration of the total transient eddy forcing of the mean anomalous streamfunction yields a different interpretation than if only the convergence of the vorticity flux is considered.

1. Introduction

Determining the effects of transient motions on the time-mean flow is one of the continuing problems in understanding the atmospheric general circulation. Much progress has been made in documenting the three-dimensional distributions of the transient eddy fluxes of heat and vorticity (or momentum) and in understanding the importance of these fluxes to the stationary flow. The studies of Holopainen (1978), Lau (1979) and Holopainen and Oort (1981), for example, indicate that transient eddy heat fluxes tend to destroy the zonally asymmetric component of the mean temperature field, while vorticity fluxes by the eddies are commonly upgradient and help to maintain the subtropical jet streams. These papers, as well as most earlier investigations, treat the local effects of transient heat fluxes independently from those of eddy vorticity forcing.

The goal of discerning the total impact of the eddies on the time-mean flow has led to a large volume of theory. For zonal mean flows, the Eliassen–Palm (E–P) flux (Andrews and McIntyre 1976; Edmon et al. 1980) provides a diagnostic of the net feedback of eddies onto the mean flow that has been widely utilized. Trenberth (1986) developed a localized E–P flux that is related to the *E*-vector of Hoskins et al. (1983) and other developments by Plumb (1985, 1986). When averaged over the middle troposphere, Holopainen et al. (1982) found that forcing by eddy heat transports dominates over eddy vorticity forcing so that the net effect is a dissipative influence on Northern Hemisphere (NH) winter stationary waves. Lau and Holopainen (1984) and Lau and Nath (1991) showed that lower-tropospheric eddy heat and vorticity fluxes reinforce one another and force cyclonic geopotential height tendencies poleward of the NH storm track axes and anticyclonic tendencies equatorward. In the upper troposphere, the polarity of the geopotential tendencies associated with eddy heat fluxes is reversed and values are considerably smaller than the tendencies associated with eddy vorticity fluxes.

Given the established importance of the transient eddy heat and momentum fluxes to the maintenance of the climatological standing waves, it is also important to establish the relationship between synoptic-scale ed-

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dies and large-scale, low-frequency waves. The modeling study of Cai and Mak (1990) demonstrated a symbiotic relationship in which low-frequency, planetary-scale waves gain energy barotropically from transient eddies which, in turn, develop in the baroclinically favorable regions created by the internally generated, low-frequency waves. This relationship was further explored using observational data for northern winter by Cai and Van den Dool (1991, 1992). Branstator (1992) used a linear budget to show that anomalous vorticity fluxes by high-frequency transients were crucial to the maintenance of recurrent low-frequency perturbation patterns in a general circulation model (GCM) simulation. His results supported the observational studies of Lau (1988), Metz (1989), and Nakamura and Wallace (1990), among others, and suggest that a substantial amount of the low-frequency variability of the midlatitude atmospheric circulation is attributable to the forcing of high-frequency eddies.

The causal relationships between low-frequency and high-frequency variability are difficult to establish because of nonlinear feedbacks. Recent work by Branstator (1995) suggests that anomalous fluxes by high-frequency eddies do not occur at random but, rather, the stormtracks are organized by the large-scale flow anomalies. Furthermore, he shows that some large-scale patterns organize storm track activity so that there is a positive feedback with high-frequency momentum fluxes while other patterns do not induce such a positive feedback. This might help explain the existence of prominent recurring low-frequency, large-scale circulation anomalies in both models and observations, and it points to a strong two-way interaction between low-frequency flow anomalies and high-frequency transients.

In assessing the impact of high-frequency eddy vorticity fluxes on either the climatological flow or large-scale, low-frequency perturbations, it has been common to make simplifying assumptions, such as only considering the horizontal flow or, in particular, only the vorticity flux convergence by the rotational component of the transient flow. The validity of these assumptions was questioned by Hoskins and Sardeshmukh (1987, hereafter HS) in a study of the total time-mean vorticity budget for an upper-tropospheric level for the mean northern winter season December 1983–February 1984. They showed that the convergence of the vorticity flux by the transient divergent flow is comparable on large scales with the equivalent rotational term, and they argued that it should not be neglected in either diagnostic studies or in parameterizations of transient eddies in global-scale climate models. A purpose of the present paper is to reexamine the results of HS using ensemble mean eddy statistics for the northern winters of November–March 1980/81–1990/91. Their primary result will be shown to be robust for the longer-term mean; namely, an analysis of transient effects should include the divergent motions when planetary scales of motion are considered. Moreover, it will be shown that the sum of the transient vertical advection and the twisting term is also important to the forcing of the mean streamfunction, and the sum of these terms primarily offsets the forcing by the vorticity flux convergence associated with the divergent flow. The result is that the convergence of the eddy vorticity flux by the total transient flow is not a good approximation to the total forcing of the long-term mean streamfunction by the high-frequency eddies.

The total direct transient eddy forcing of the vorticity balance in the upper-troposphere will also be examined for two cases of anomalous large-scale flow. One case involves changes in the mean state and the associated changes in the storm tracks and anomalous transient eddy fluxes of vorticity associated with extremes of the North Atlantic Oscillation, a primary mode of atmospheric interannual variability (e.g., Wallace and Gutzler 1981). The second case relates to the influence of high-frequency eddy vorticity fluxes on a pronounced low-frequency circulation anomaly over the North Pacific that persisted for more than a decade (Trenberth 1990a; Trenberth and Hurrell 1994).

2. Methodology

a. Data

The European Centre for Medium-Range Weather Forecasts (ECMWF) global analyses used in this study are described by Trenberth (1992). They are twice daily (0000 and 1200 UTC) and have been processed at T42 resolution with spectral transform techniques to exactly compute vorticity, divergence, and the rotational and divergent wind components. The transient statistics were obtained by applying a bandpass filter that retains fluctuations between 2 and 8 days. Therefore, only the effects of synoptic timescale transients will be considered. The bandpass filter, its response function, and the percent variance retained for several fields are presented in Trenberth (1991). Changes in analysis procedures at ECMWF since 1979 have introduced spurious changes in several fields, although the effects are mostly small on the filtered data, especially in the extratropics (Trenberth 1991), and are of no consequence to the results presented here.

The results of HS were from ECMWF analyses on a global 5° latitude–longitude grid for the period December 1983–February 1984. They examined the transient vorticity forcing by eddies for all subseasonal timescales in addition to synoptic timescale transients, which were isolated with a 31-point Lanczos filter (Duchon 1979) applied to daily 1200 UTC data. This filter passes periods from 2.16 to 6.15 days. Wallace et al. (1988) show that the properties of high-frequency transients are fairly insensitive to the use of different high-pass filters, so using the 2–8-day filter of Trenberth (1991) should present no serious comparison problems.

b. The vorticity equation

Following HS, the direct transient eddy forcing on the seasonal-mean vertical component of the vorticity equation in pressure coordinates is

$$\left.\frac{\partial\bar{\zeta}}{\partial t}\right|_e = -\nabla\cdot\overline{\mathbf{v}'\zeta'} - \mathbf{k}\cdot\nabla\times\overline{\omega'\frac{\partial\mathbf{v}'}{\partial p}}, \quad (1)$$

where frictional and small-scale processes have been ignored, and the last term on the right-hand side of (1) is the sum of the vertical advection and twisting terms.

Using the anisotropic eddy correlation components discussed in Hoskins et al. (1983), the horizontal vorticity flux by transient eddies can be partitioned as

$$-\nabla\cdot\overline{\mathbf{v}'\zeta'} = G(M, N) + \mathbf{k}\cdot\nabla\times\overline{\mathbf{v}'\delta'}, \quad (2)$$

so that (1) becomes

$$\left.\frac{\partial\bar{\zeta}}{\partial t}\right|_e = G(M, N) + \mathbf{k}\cdot\nabla\times\overline{\mathbf{v}'\delta'} - \mathbf{k}\cdot\nabla\times\overline{\omega'\frac{\partial\mathbf{v}'}{\partial p}} \quad (3)$$

or

$$\left.\frac{\partial\bar{\zeta}}{\partial t}\right|_e = G(M, N) - \mathbf{k}\cdot\nabla\times\frac{\partial}{\partial p}\overline{\omega'\mathbf{v}'}, \quad (4)$$

where

$$G(M, N) = 2M_{xy} + N_{yy} - N_{xx}$$

and

$$M = \frac{1}{2}(\overline{u'^2} - \overline{v'^2}), \quad N = \overline{u'v'}.$$

In the above equations, ζ is the absolute vorticity, δ is the divergence, overbars represent the mean flow, and primes represent the bandpass time-filtered transient flow. All other symbols have their usual meteorological meanings.

To compare the importance of the flux of vorticity by the divergent component of the transient flow to the flux by the rotational component, it is useful to split the convergence of the transient horizontal vorticity flux into rotational (ψ) and divergent (χ) components:

$$-\nabla\cdot\overline{\mathbf{v}'\zeta'} = -\nabla\cdot\overline{\mathbf{v}'_\psi\zeta'} - \nabla\cdot\overline{\mathbf{v}'_\chi\zeta'} \quad (5)$$

or

$$-\nabla\cdot\overline{\mathbf{v}'\zeta'} = G(M_\psi, N_\psi) - \nabla\cdot\overline{\mathbf{v}'_\chi\zeta'}. \quad (6)$$

3. Maintenance of the mean flow

One point made by HS was that the vorticity flux by the transient divergent flow $\overline{\mathbf{v}'_\chi\zeta'}$ was of similar magnitude to that by the rotational flow $\overline{\mathbf{v}'_\psi\zeta'}$ for the winter of 1983/84. Figure 1 shows these fields at 300 mb for the November–March 1980/81–90/91 ensemble mean. The 300-mb level was selected since it resides within the troposphere at all latitudes. In addition, the divergent flow is strong at this level, and the results presented hereafter do not differ significantly from those obtained at other upper-tropospheric levels. Many of the same features discussed in HS are apparent in Fig. 1 as well. In particular, the two vorticity fluxes are of nearly equal magnitude, and the vorticity flux by the transient divergent flow shows a very coherent structure with maximum amplitude in the storm track regions of both hemispheres. The result of this coherent structure, however, is that the magnitude of the convergence of this term is small compared to that of $\overline{\mathbf{v}'_\psi\zeta'}$ (Fig. 2). The relevant point from Fig. 2 is that the traditional assumptions of the dominance of the rotational component made in assessing the importance of these terms in forcing the mean flow seem reasonable. The pattern that emerges is one of cyclonic forcing poleward and anticyclonic forcing equatorward of the main storm tracks in both hemispheres. The convergence of $\overline{\mathbf{v}'_\chi\zeta'}$ is strongest in the western Pacific and Atlantic in the NH and primarily works to offset the anticyclonic forcing from the transient rotational flow on the equatorward side of the storm tracks.

If the focus is changed to planetary scales by examining the equivalent forcing of the mean streamfunction, the importance of the transient divergent vorticity flux becomes more apparent. This is shown in Fig. 3, which is simply the inverse Laplacian of the fields in Fig. 2. The larger-scale features emphasized by the streamfunction perspective are probably the ones of greater significance on monthly and longer timescales relative to the smaller spatial scales brought out by the vorticity view presented in Fig. 2. For the November–March northern winter season, the main effect of $-\nabla\cdot\overline{\mathbf{v}'_\chi\zeta'}$ on the mean streamfunction is a strong cyclonic forcing in the vicinity of the East Asian jet, with weaker cyclonic forcings on the poleward side of the North Atlantic jet and near the Atlantic and Indian Ocean portions of the Southern Hemisphere (SH) storm track. The advection by the rotational component of the flow, or equivalently $-\nabla\cdot\overline{\mathbf{v}'_\psi\zeta'}$, makes up a smaller part of the total advection in the North Pacific but, on planetary scales, results in a clear pattern of cyclonic forcing on the poleward side and anticyclonic forcing on the equatorward side of the main storm tracks, as in Fig. 2. The anticyclonic forcing in the North Pacific is offset by the strong cyclonic forcing of the $-\nabla\cdot\overline{\mathbf{v}'_\chi\zeta'}$ term, however, so that the total effect of the convergence of the transient horizontal vorticity flux is a cyclonic tendency across this region.

The importance of the divergent flow is also evident when one considers (2), which makes use of the anisotropic eddy correlation components of Hoskins et al. (1983). When the curl of the eddy divergence flux is compared to $G(M, N)$ (not shown), it again appears as though the assumption of horizontal nondivergent flow is good since $G(M, N)$ is the dominant term.

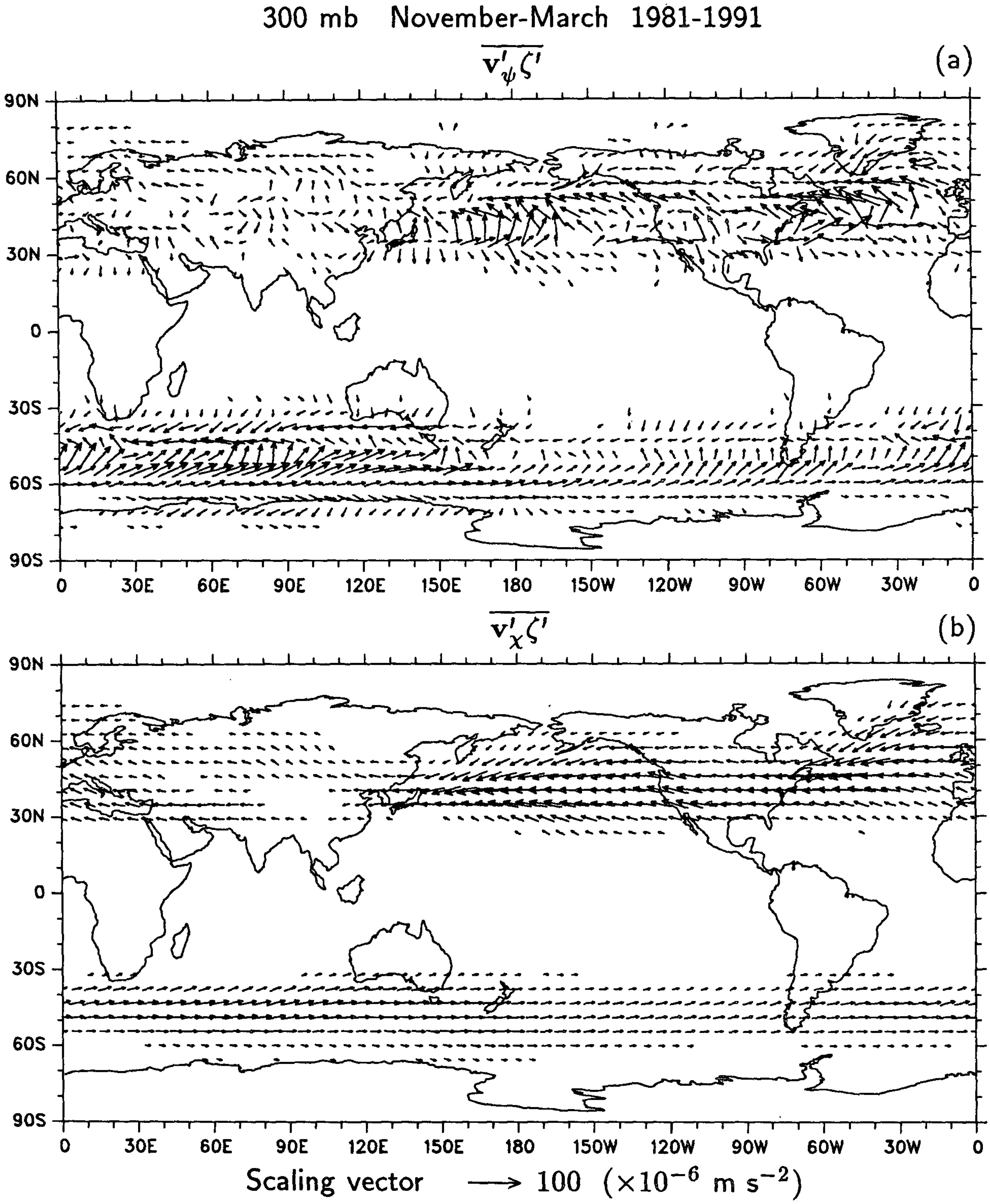

FIG. 1. Vorticity fluxes at 300 mb for November–March 1980/81–1990/91 associated with (a) the transient rotational flow and (b) the transient divergent flow. The scaling arrow is the same for both panels (100×10^{-6} m s^{-2}). The results have been truncated to T21 resolution.

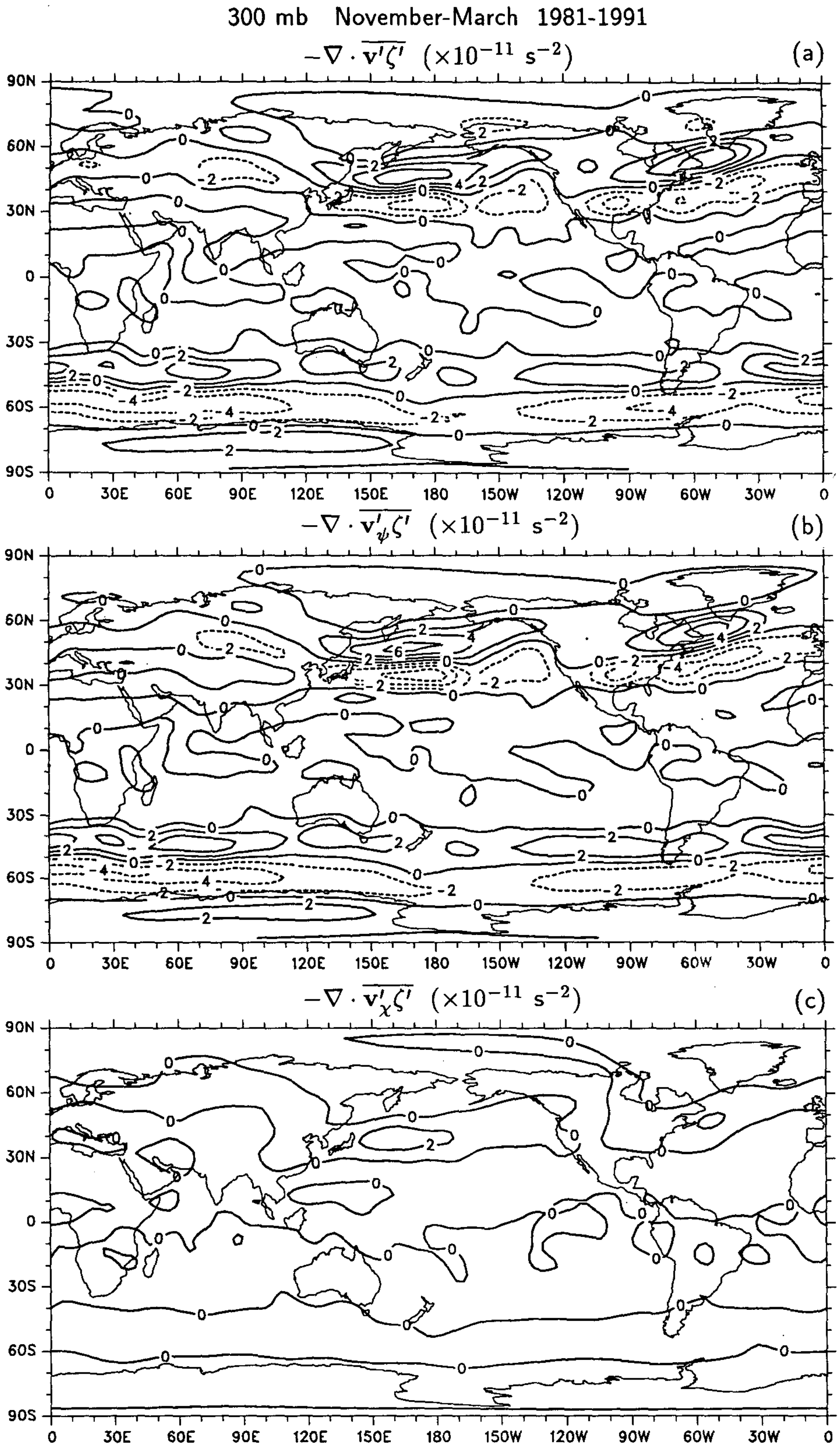

FIG. 2. Convergence of the vorticity flux by (a) the total transient flow, (b) the transient rotational flow, and (c) the transient divergent flow. The contour increment is 2×10^{-11} s^{-2}. Negative values are dashed, and results are truncated to T21 resolution.

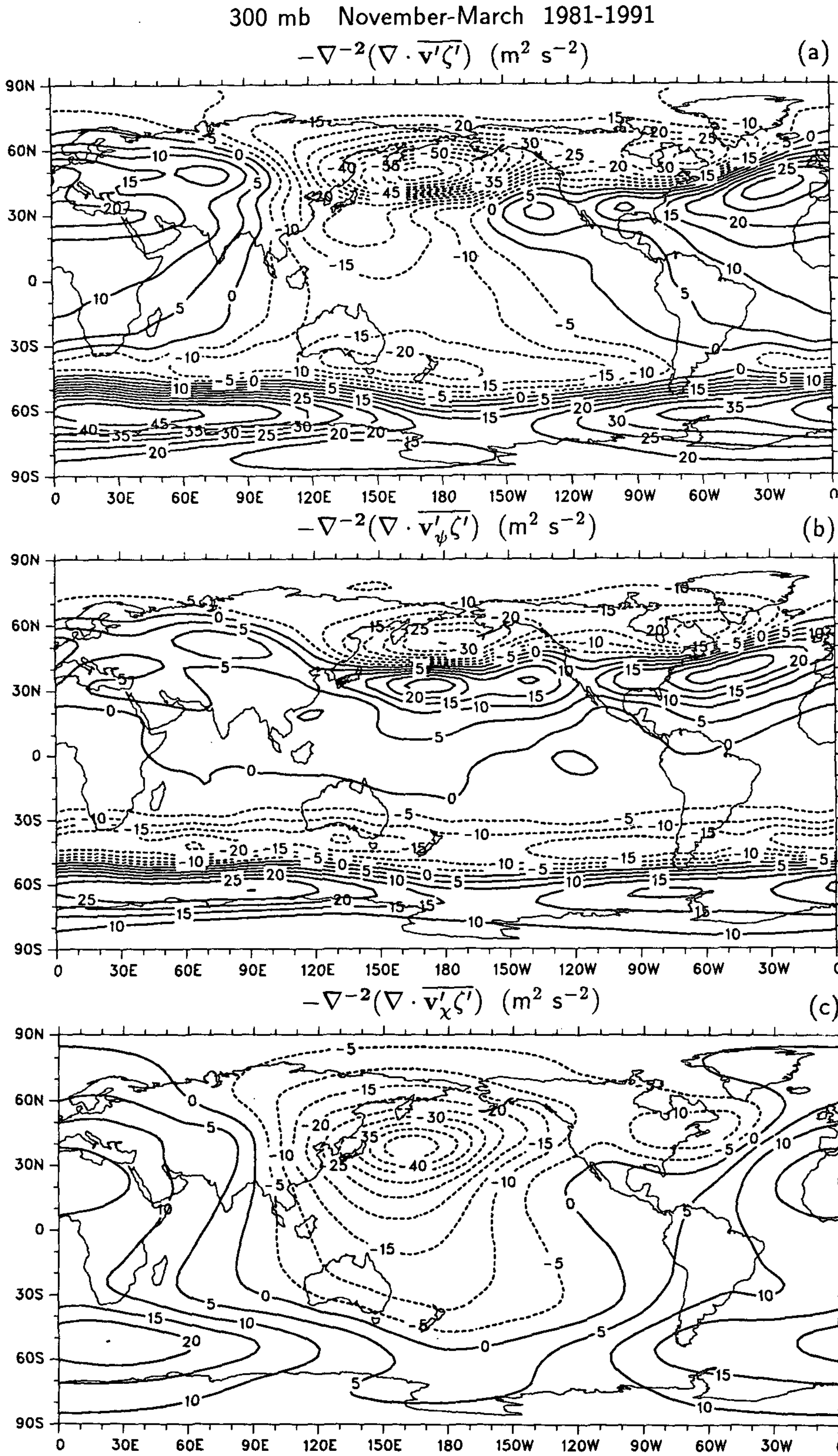

FIG. 3. The streamfunction forcing associated with the convergence of the vorticity flux by (a) the total transient flow, (b) the transient rotational flow, and (c) the transient divergent flow. The contour increment is 5 m^2 s^{-2}. Negative values are dashed, and results are truncated to T21 resolution.

However, the assumption is poor when considering the forcing of the mean streamfunction (Fig. 4). The curl of the eddy divergence flux is responsible for a large cyclonic forcing on the poleward side of the main storm tracks (Fig. 4b), while $G(M, N)$ acts to force anticyclonic tendencies on the equatorward sides and cyclonic tendencies on the poleward sides of the storm tracks in both hemispheres (Fig. 4a). The similarity between the streamfunction forcing by the curl of the eddy divergence flux and $-\nabla\cdot\overline{\mathbf{v}'_\chi\zeta'}$ in Fig. 3c arises from (2) and (6) because the eddy correlation tensor components M and N are well approximated by M_ψ and N_ψ computed from the rotational flow alone (Hoskins et al. 1983; HS). Indeed, for the November–March 1980/81–1990/91 ensemble mean, $G(M_\chi, N_\chi)$ is an order of magnitude smaller than $G(M_\psi, N_\psi)$ (not shown). The salient point is that both terms on the right-hand side of (2) contribute significantly to the forcing of the rotational flow by the transient eddy horizontal vorticity flux shown in Fig. 3a.

The size of the eddy divergence flux in (2) is an indication that $-\mathbf{k}\cdot\nabla\times\overline{\omega'(\partial\mathbf{v}'/\partial p)}$ in (1) and (3) needs to be considered more carefully to evaluate the total eddy forcing of the mean flow. This term has traditionally been assumed to be relatively small, and indeed it is when the mean vorticity budget is examined. When the forcing of the mean streamfunction is considered, however, this is not the case and $-\mathbf{k}\cdot\nabla\times\overline{\omega'(\partial\mathbf{v}'/\partial p)}$ is large and primarily works to reduce the magnitude of the cyclonic forcing of $-\nabla\cdot\overline{\mathbf{v}'\zeta'}$ in (1). In fact, it is roughly equal but of opposite sign to the curl of the eddy divergence flux shown in Fig. 4b. The residual between these two large terms of roughly opposite sign in (3) is the vertical momentum-flux convergence in (4), $-\mathbf{k}\cdot\nabla\times(\partial/\partial p)\overline{\omega'\mathbf{v}'}$. This quantity is largest in the East Asian jet region and represents a cyclonic forcing in excess of 20 $\mathrm{m^2\,s^{-2}}$ (Fig. 4c). A latitude–height cross section of $-\mathbf{k}\cdot\nabla\times\overline{\omega'\mathbf{v}'}$ zonally averaged across the Pacific storm track from 140°E to 170°E is shown in Fig. 5. This quantity is a maximum from about 850 up to 500 mb, so that the vertical derivative exhibits a maximum in the lower troposphere with a secondary maximum of opposite sign in the upper troposphere.

The above results show that the net effect of the transient eddies on the time-mean streamfunction (Fig. 4d) is well approximated by $G(M, N)$ (Fig. 4a), but in the upper troposphere $G(M, N)$ alone would underestimate the cyclonic forcing with the largest error in the North Pacific. Perhaps more importantly, the forcing of the mean streamfunction by the convergence of the vorticity flux by the total transient flow in Fig. 3a is not a good approximation to the net effect of the eddies (Fig. 4d), although it is frequently used in this capacity. In particular, $\nabla^{-2}(-\nabla\cdot\overline{\mathbf{v}'\zeta'})$ significantly overestimates the cyclonic forcing on the poleward side of the storm tracks in both hemispheres and underestimates the anticyclonic forcing on the equatorward sides. In other words, $-\mathbf{k}\cdot\nabla\times\overline{\omega'(\partial\mathbf{v}'/\partial p)}$ in (1) makes a significant contribution to the transient eddy vorticity forcing of the climatological mean flow on planetary scales. Moreover, since on large scales $-\nabla\cdot\overline{\mathbf{v}'_\chi\zeta'}\approx\mathbf{k}\cdot\nabla\times\overline{\mathbf{v}'\delta'}$ (Figs. 3c and 4b), it follows that $-\nabla\cdot\overline{\mathbf{v}'_\chi\zeta'}\approx\mathbf{k}\cdot\nabla\times\overline{\omega'(\partial\mathbf{v}'/\partial p)}$, where $-\mathbf{k}\cdot\nabla\times(\partial/\partial p)\overline{\omega'\mathbf{v}'}$ is small (Fig. 4c). The result is that the total eddy forcing of the long-term mean November–March streamfunction is better given by $-\nabla\cdot\overline{\mathbf{v}'_\psi\zeta'}$ or $G(M, N)$ than by the convergence of the vorticity flux by the total transient flow, $-\nabla\cdot\overline{\mathbf{v}'\zeta'}$.

4. Maintenance of an anomalous flow in the North Pacific

In a recent paper, Trenberth and Hurrell (1994, hereafter TH) documented significant variations on decadal timescales in the North Pacific that involved both atmosphere and ocean. They focused particular attention on a change that began about 1976 which, within the atmosphere, corresponded to a deeper and eastward shifted Aleutian Low pressure system in the winter half year, November through March. The link of the North Pacific circulation to the Tropics and the Southern Oscillation was examined from a causal perspective, as were the relationships between the changes in the mean state and the associated changes in the storm tracks and anomalous transient eddy fluxes of heat and vorticity.

Because high-quality global analyses do not exist prior to 1979, TH could not document the exact changes in the storm tracks and the effects of the changes in the transient eddy fluxes that occurred with the change in 1976. To draw an analogy to the low-frequency change, TH made use of the large interannual variability in the North Pacific and contrasted winters in which the surface pressure of the Aleutian low was close to a climatological value (a "normal" winter) against those in which the Aleutian low was anomalously deep (a "low" winter). A simple index (referred to as the North Pacific or NP index) of the area-weighted mean sea level pressure over the region 30° to 65°N, 160°E to 140°W was defined to measure the interannual variations (see their Figs. 6 and 7).

In the following discussion, the low NP index winter composite is the average November through March bandpassed ECMWF analyses for the winters 1980/81, 1982/83, 1983/84, 1985/86, and 1986/87 (the given year corresponds to January). The normal NP index composite is the average of 1981/82, 1984/85, 1988/89, 1989/90, and 1990/91 winters. The anomalous streamfunction, low-normal NP values, at 300 mb is shown in Fig. 6. A global map is used to illustrate the extension of the anomalous wavetrain from the Tropics across the North Pacific and downstream over the United States. The dominant feature is a large cyclonic circulation centered at ~45°N in the North Pacific, and

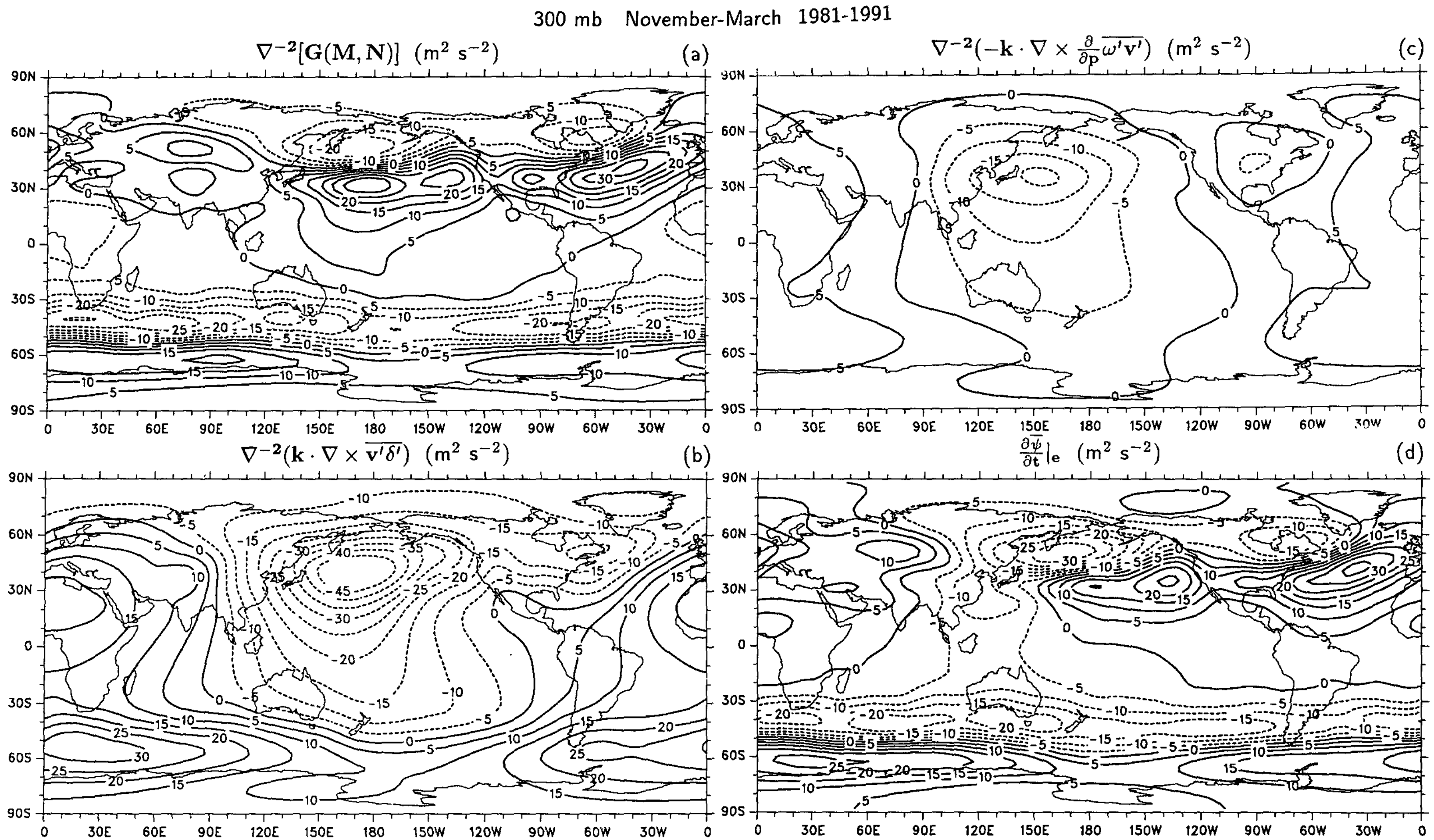

FIG. 4. The streamfunction forcing associated with (a) $G(M, N)$, (b) the curl of the transient divergence flux, and (c) the transient vertical momentum-flux convergence. The total transient forcing is shown in (d), which is the sum of (a) and (c). The contour increment is 5 $\mathrm{m^2\ s^{-2}}$. Negative values are dashed, and results are truncated to T21 resolution.

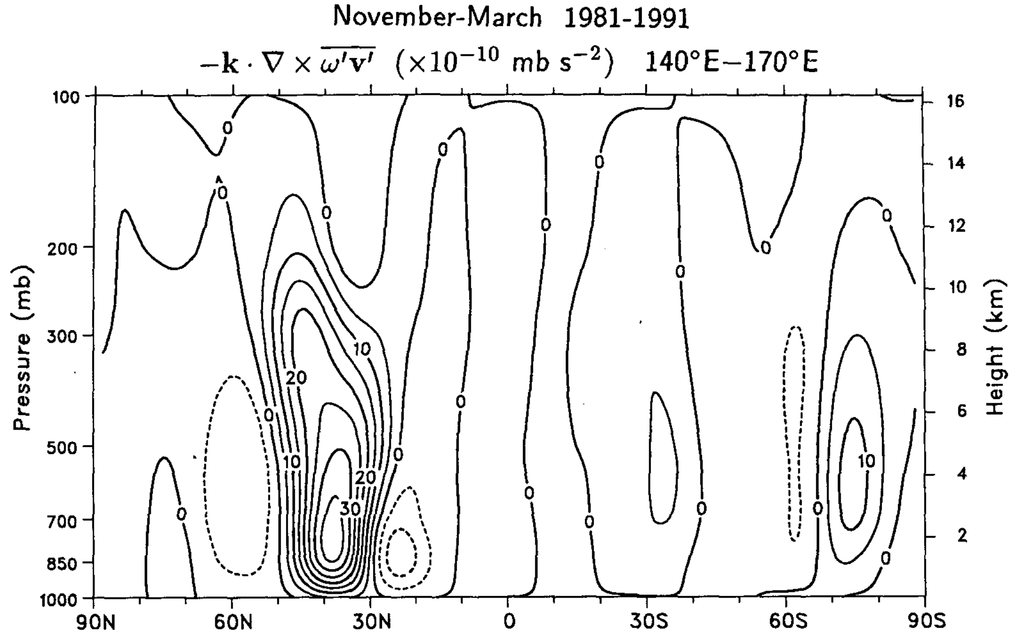

FIG. 5. Latitude–height cross section of $-\mathbf{k}\cdot\nabla\times\overline{\omega'\mathbf{v}'}$ zonally averaged across the Pacific from 140° to 170°E. The contour increment is 5×10^{-10} mb s^{-2}. Negative values are dashed, and results are truncated to T21 resolution.

an anticyclonic circulation of equal magnitude ~2500 km directly to the south. The result is a westerly wind anomaly greater than 15 m s^{-1} near 30°N, with anomalous easterlies to the north and south. During low NP years, TH showed that a shift occurs in the position of the mean storm tracks with enhanced activity across the North Pacific south of 40°N and downstream over the southern United States, and reduced activity farther north (their Figs. 13 and 14). This systematic change allows the possibility that changes in the transient heat and vorticity fluxes may have helped maintain the anomalous mean circulation over the North Pacific.

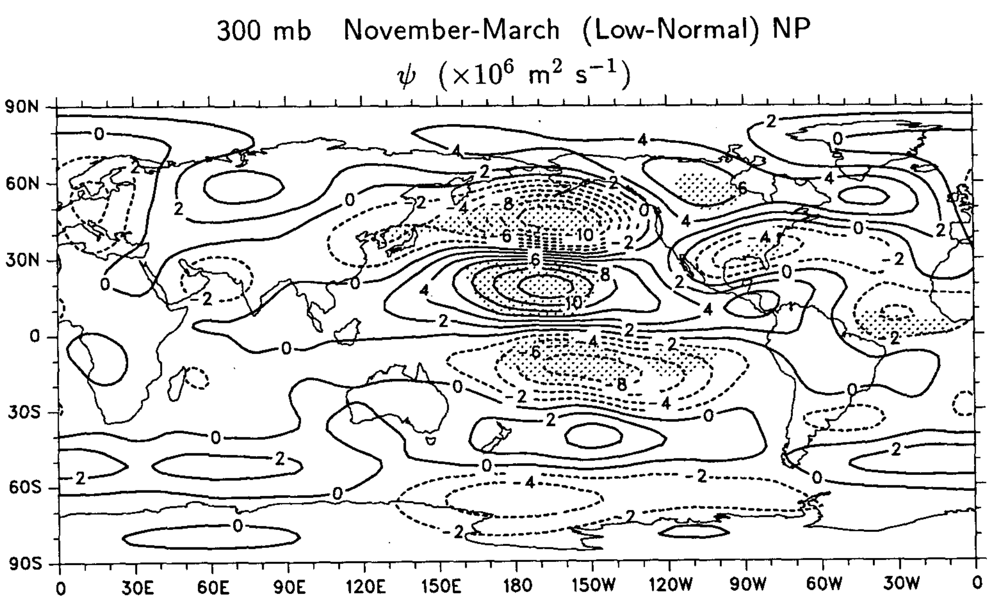

FIG. 6. Anomalous 300-mb streamfunction for the low–normal NP composite. The contour increment is 2×10^6 m^2 s^{-1}. Negative values are dashed, and results are truncated to T21 resolution. Values significantly different from zero at the 5% level using a t test are stippled.

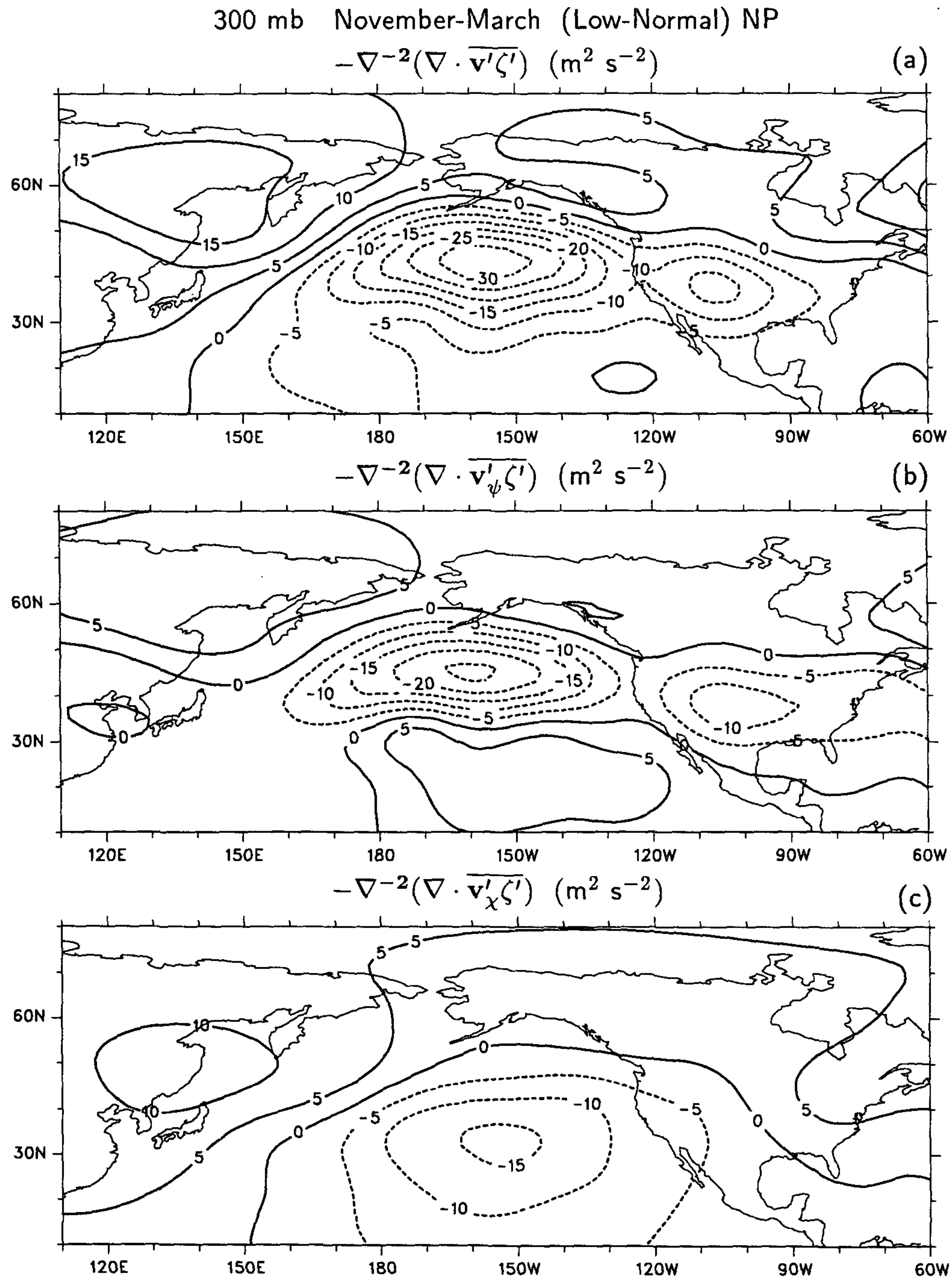

FIG. 7. The streamfunction forcing, low–normal NP years, associated with the convergence of the vorticity flux by (a) the total transient flow, (b) the transient rotational flow, and (c) the transient divergent flow. The contour increment is 5 m^2 s^{-2}. Negative values are dashed, and results are truncated to T21 resolution.

Trenberth and Hurrell (1994) examined this question by computing localized E–P fluxes following the formulation of Trenberth (1986). Since vorticity fluxes dominate over heat fluxes in the upper troposphere, the change in the horizontal E–P flux component is dominant and it proves easier to summarize the results by examining the transient eddy forcing of the mean streamfunction.

The quantity $\nabla^{-2}(-\nabla \cdot \overline{\mathbf{v}'\zeta'})$ at 300 mb for the low–normal NP composite is presented in Fig. 7a. The clear result is that transient vorticity fluxes systematically reinforce and help to maintain the anomalous upper-tropospheric rotational flow for a low NP index. Similar results from linear model analyses of the role of transient eddy vorticity fluxes to the maintenance of anomalous extratropical wavetrains over the Pacific have been noted in many studies (e.g., Kok and Opsteegh 1985; Held and Kang 1987; Held et al. 1989; Ting and Hoerling 1993; Hoerling and Ting 1994). The largest contribution to this pattern for the low–normal NP composite arises from the convergence of the vorticity flux by the transient rotational flow (Fig. 7b). The convergence of the vorticity flux by the divergent flow is significant, however, and it contributes to the overall cyclonic tendency in the region (Fig. 7c).

The net effect of all eddy terms in the vorticity equation on the anomalous streamfunction is shown in Fig. 8. This field is dominated by the anisotropic eddy correlation component $G(M, N)$ in (4), which exhibits a dipole structure of cyclonic forcing north of 40°N and anticyclonic forcing southeast of there in the North Pacific. The vertical momentum flux convergence term in (4) (not shown) contributes a cyclonic tendency such that neglect of this term would result in approximately a 20% underestimation of the maximum cyclonic forcing near 45°N, 165°W. An important point is that the streamfunction forcing by $-\mathbf{k}\cdot\nabla \times \overline{\omega'(\partial \mathbf{v}'/\partial p)}$ in (1) (not shown) primarily works to counteract the forcing by $-\nabla\cdot\overline{\mathbf{v}'_\chi \zeta'}$ (Fig. 7c) so that, compared to the convergence of the vorticity flux by the total anomalous transient flow in Fig. 7a, the net vorticity forcing (Fig. 8) shows a slight northwestward displacement of the maximum cyclonic circulation and a stronger subtropical anticyclonic forcing. The anomalous cyclonic forcing in Fig. 8, acting unopposed, would spin up streamfunction anomalies of the observed magnitude in ~6 days, compared to a faster timescale of ~4 days implied by the convergence of the vorticity flux alone (Fig. 7a). The anticyclonic forcing in Fig. 8 is in the same region noted by TH as having statistically significant positive anomalies of outgoing longwave radiation (see their Fig. 18), a signature associated with large-scale overturning. Moreover, the overall picture that emerges agrees with the conclusion of Held et al. (1989) that significant central Pacific subtropical convergence anomalies in the large-scale mean flow are strongly related to anomalous transient vorticity fluxes, which then can feedback and reinforce the anomalous extratropical wavetrain.

5. Maintenance of an anomalous flow in the North Atlantic

One of the primary modes of atmospheric interannual variability is the North Atlantic Oscillation (NAO) (e.g., van Loon and Rogers 1978; Wallace and Gutzler 1981; Barnston and Livezey 1987; Kushnir and Wallace 1989). The NAO corresponds to a meridional oscillation in atmospheric mass with centers of action near the Icelandic low and the Azores high. Although it is evident throughout the year, it is most pronounced during NH winter (Barnston and Livezey 1987) and accounts for approximately one-third of the total variance of the sea level pressure field over the North Atlantic Ocean basin (Deser and Blackmon 1993).

A time series of a simple index of the NAO reveals considerable variability over the past century on interannual and longer timescales (Hurrell and van Loon 1995). Such variations over the Atlantic relate to distinct patterns of coupled ocean–atmosphere relationships (Cayan 1992; Deser and Blackmon 1993; Kushnir 1994) as well as to significant changes and trends in regional temperatures and precipitation (e.g., van Loon and Williams 1976; van Loon and Rogers 1978; Trenberth 1990b). The NAO index is defined as the

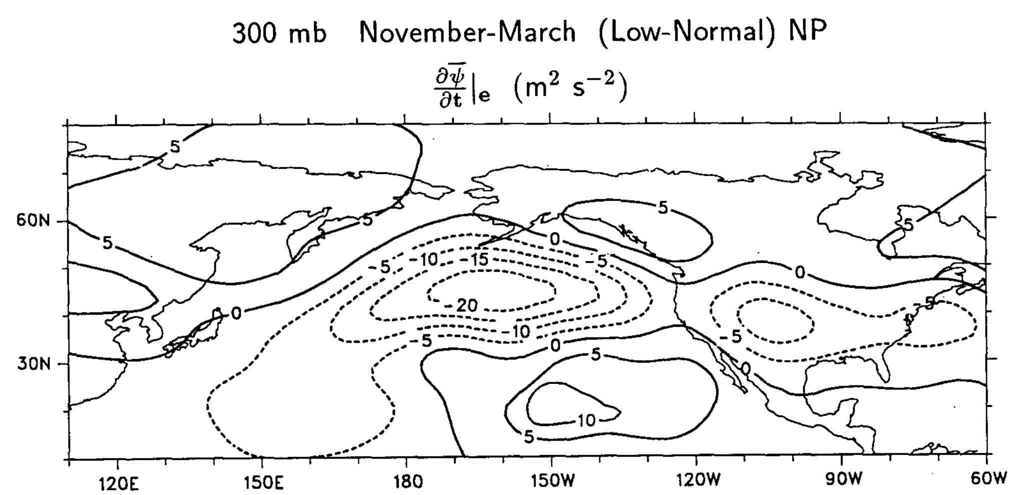

FIG. 8. The total transient eddy forcing of the streamfunction, low–normal NP years. The contour increment is 5 m^2 s^{-2}. Negative values are dashed, and results are truncated to T21 resolution.

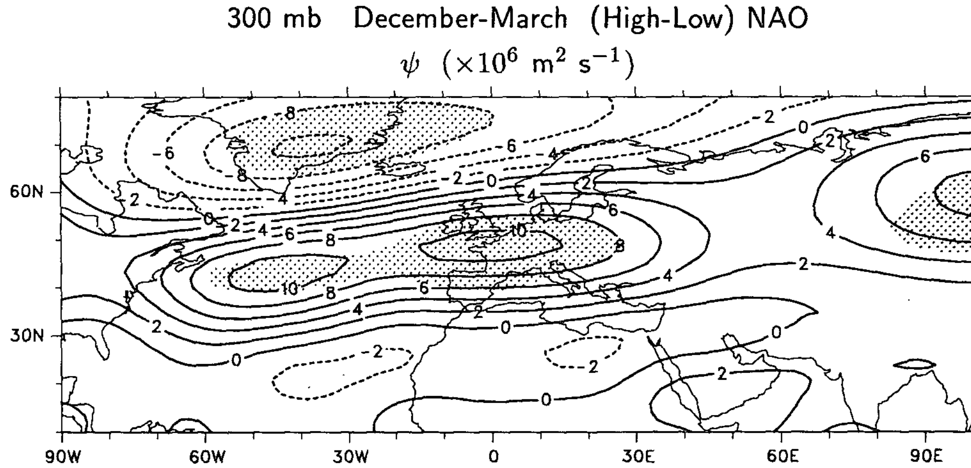

FIG. 9. Anomalous 300-mb streamfunction for the high–low NAO composite. The contour increment is 2×10^6 m^2 s^{-1}. Negative values are dashed and results are truncated to T21 resolution. Values significantly different from zero at the 5% level using a t test are stippled.

difference in normalized December through March[1] sea level pressure anomalies between Ponta Delgadas, Azores, and Akureyri, Iceland (e.g., Rogers 1984). The winter pressure anomalies at each station were normalized by dividing each seasonal pressure by the long-term mean (1894–1993) standard deviation. More complex indices of the NAO can be defined, but Wallace and Gutzler (1981; their Fig. 7a) confirm that the Azores and Iceland are the main centers of teleconnection in pressure associated with the oscillation.

Compositing the ECMWF eddy statistics over years in which the NAO index (defined as above) is greater than 1.0 and those when the index is less than -1.0 (high index years minus low index years) indicates a clear shift in storm track activity (not shown) with statistically significant enhanced variance over the North Atlantic and northern Europe and reduced activity over the subtropical Atlantic. The high-NAO index winter composite in the following figures is the average December through March bandpassed analyses for the winters of 1980/81, 1982/83, 1983/84, 1988/89, and 1989/90, while the low NAO index winters include 1978/79, 1979/80, 1984/85, and 1986/87.

The anomalous 300-mb streamfunction for high–low NAO index winters (Fig. 9) consists of a pronounced anticyclonic circulation stretching across the Atlantic Ocean and Europe between about 35° and 55°N and a strong cyclonic circulation centered over Greenland and Iceland. The forcing of this pattern through the convergence of the vorticity flux by anomalous transients is presented in Fig. 10. As in the North Pacific case study, the eddies generally reinforce and help to maintain the anomalous upper-tropospheric rotational flow for a high NAO index. The convergence of the vorticity flux by the transient rotational flow again makes the largest contribution (Fig. 10b), although $\nabla^{-2}(-\nabla\cdot\overline{\mathbf{v}'_\chi\zeta'})$ is locally significant and primarily forces a cyclonic circulation over the Norwegian and North Seas (Fig. 10c).

The net effect of all transient eddy terms in the vorticity equation on the anomalous streamfunction is shown in Fig. 11. There are significant differences from the forcing by $\nabla^{-2}(-\nabla\cdot\overline{\mathbf{v}'\zeta'})$ (Fig. 10a), which is the lone term that is most frequently used to represent the effect of the transient eddies. In particular, the net cyclonic forcing in Fig. 11 is less and is further west than in Fig. 10a and, as for the North Pacific case, would spin up streamfunction anomalies of the observed magnitude in ~6 days if acting alone. The anticyclonic forcing near 45°N is also stronger and is slightly more extensive when all transient eddy terms are considered. In general, the net streamfunction forcing corresponds well to the anomalous circulation centers evident in Fig. 9. The pattern in Fig. 11 is dominated by the anisotropic eddy correlation component $G(M, N)$, with the vertical momentum flux convergence term (not shown) making a smaller contribution than for the North Pacific case. Just as for the climatological standing waves and the North Pacific case, the importance of $-\mathbf{k}\cdot\nabla \times \overline{\omega'(\partial\mathbf{v}'/\partial p)}$ in (1) to the forcing of the anomalous streamfunction is evident in that the total transient eddy forcing is better approximated by $-\nabla\cdot\overline{\mathbf{v}'_\psi\zeta'}$ than by $-\nabla\cdot\overline{\mathbf{v}'\zeta'}$.

6. Summary

Hoskins and Sardeshmukh (1987) examined the total time-mean vorticity budget at 250 mb for the mean

[1] Winter is better defined as the four-month season in the Atlantic because November is more typical of a transitional month (see Hurrell and van Loon 1995).

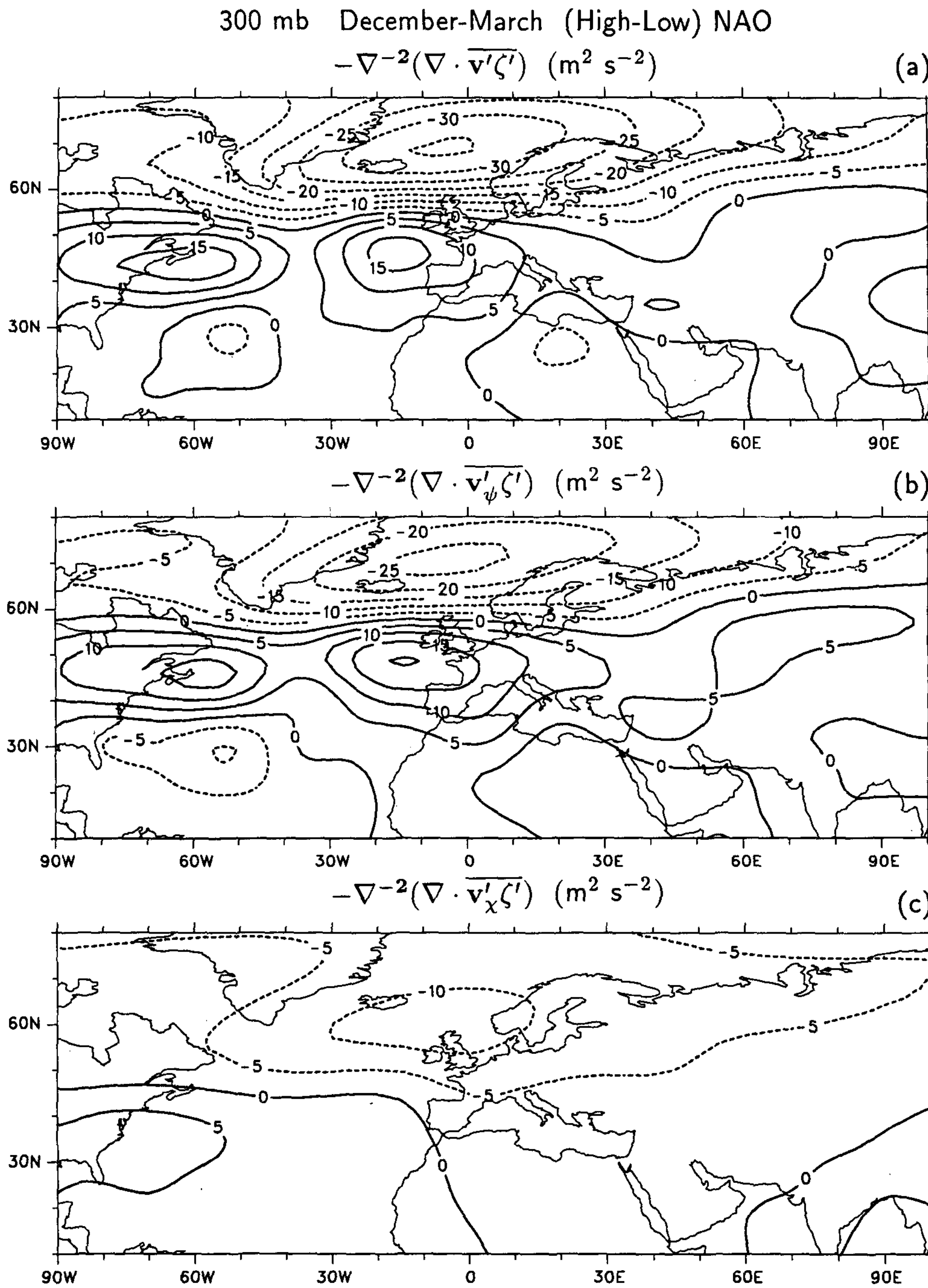

FIG. 10. The streamfunction forcing, high–low NAO years, associated with the convergence of the vorticity flux by (a) the total transient flow, (b) the transient rotational flow, and (c) the transient divergent flow. The contour increment is 5 m^2 s^{-2}. Negative values are dashed, and results are truncated to T21 resolution.

northern winter season December 1983–February 1984. The principal finding of their work was that terms associated with the horizontally divergent transient motions are important, especially when planetary scales of motion are considered. In particular, the convergence of the vorticity flux associated with the divergent motion is comparable with the equivalent rotational term when the forcing of the mean streamfunction is examined. This result was shown to be robust in this paper using 11 years of bandpass-filtered analyses from

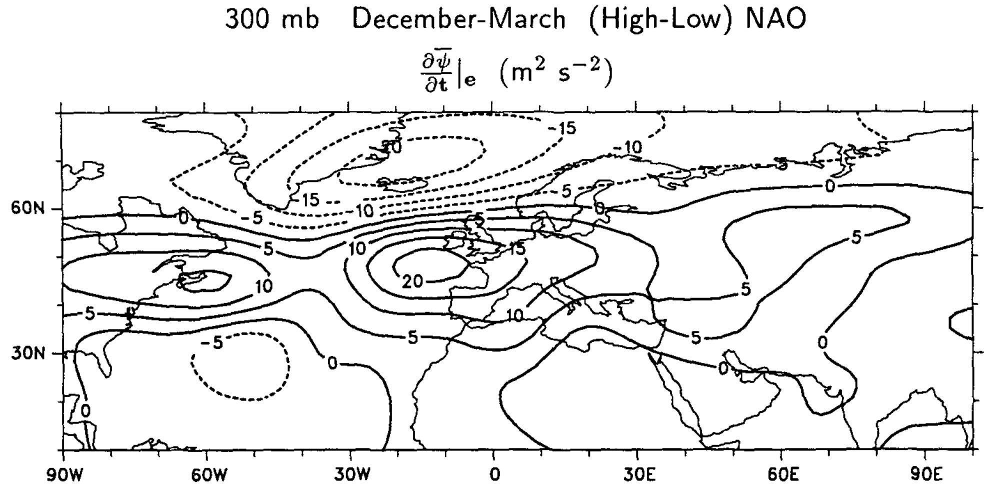

FIG. 11. The total transient eddy forcing of the streamfunction, high–low NAO years. The contour increment is 5 $m^2\ s^{-2}$. Negative values are dashed, and results are truncated to T21 resolution.

ECMWF that retain fluctuations between 2 and 8 days. For the mean northern winter season November–March 1980/81–1990/91, the streamfunction forcing associated with $-\nabla\cdot\overline{\mathbf{v}'_\chi\zeta'}$ is largest is the East Asian jet region where it produces a cyclonic tendency of nearly 45 $m^2\ s^{-2}$ (Fig. 3c). This term also produces significant, although smaller, cyclonic forcings in the North Atlantic and SH storm track regions, so that a consideration of the streamfunction forcing by $-\nabla\cdot\overline{\mathbf{v}'_\psi\zeta'}$ alone produces a substantially different interpretation of the transient forcing.

A new finding is the importance of $-\mathbf{k}\cdot\nabla\times\overline{\omega'(\partial\mathbf{v}'/\partial p)}$ in (1) and (3), which is the sum of the transient vertical advection and twisting terms. In the vertical integral when ω' is zero at the upper and lower boundaries, this term vanishes, which is one reason why it has been neglected in previous studies (e.g., Holopainen and Oort 1981). The importance at any particular level in the troposphere is less clear, however, as discussed by HS. They present maps of the curl of the eddy divergence flux, $\mathbf{k}\cdot\nabla\times\overline{\mathbf{v}'\delta'}$, and the vertical eddy momentum flux convergence, $-\mathbf{k}\cdot\nabla\times(\partial/\partial p)\overline{\omega'\mathbf{v}'}$, at 250 mb, but they conclude that the similarity of these terms confirms the negligible value of $-\mathbf{k}\cdot\nabla\times\overline{\omega'(\partial\mathbf{v}'/\partial p)}$ in the upper troposphere. This assumption appears to be untrue when the forcing of the long-term mean streamfunction is considered. In fact, $-\mathbf{k}\cdot\nabla\times\overline{\omega'(\partial\mathbf{v}'/\partial p)}$ largely balances the curl of the eddy divergence flux in (3) (Fig. 4b), leaving the forcing given by $-\mathbf{k}\cdot\nabla\times(\partial/\partial p)\overline{\omega'\mathbf{v}'}$ in Fig. 4c. The primary effect of the $-\mathbf{k}\cdot\nabla\times\overline{\omega'(\partial\mathbf{v}'/\partial p)}$ in (1), therefore, is to reduce the magnitude of the cyclonic forcing of $-\nabla\cdot\overline{\mathbf{v}'\zeta'}$ on the poleward side of the storm tracks in both hemispheres (cf. Figs. 3a and 4d). The result is that an approximation of the total transient-eddy forcing of the mean streamfunction is better given by $G(M, N)$ in (4), although the vertical eddy momentum flux convergence makes an important contribution in the vicinity of the East Asian jet. Furthermore, it is interesting to note that since $G(M, N) \approx G(M_\psi, N_\psi) = -\nabla\cdot\overline{\mathbf{v}'_\psi\zeta'}$, the total eddy forcing on planetary scales is better approximated by the convergence of the vorticity flux by the transient rotational flow than by the total transient flow, $-\nabla\cdot\overline{\mathbf{v}'\zeta'}$. This is true to the extent that the streamfunction forcing by $-\mathbf{k}\cdot\nabla\times\overline{\omega'(\partial\mathbf{v}'/\partial p)}$ balances $-\nabla\cdot\overline{\mathbf{v}'_\chi\zeta'}$ in (1).

The role of each transient eddy term in maintaining large-scale anomalous flows was also examined. In both the North Pacific and North Atlantic case studies, transient vorticity fluxes systematically reinforce and help to maintain the upper-tropospheric rotational flow anomalies. This finding agrees with the model analyses of Held et al. (1989), Lau and Nath (1990), Kushnir and Lau (1992), and Branstator (1992) among others. Just as for the climatological flow, consideration of all eddy terms in the vorticity equation yields a different result than when the transient contribution is given solely by $\nabla^{-2}(-\nabla\cdot\overline{\mathbf{v}'\zeta'})$. In particular, the convergence of the vorticity flux by the transient flow significantly overestimates the cyclonic forcing by the eddies and underestimates the induced anticyclonic tendencies. This may be an important point to consider when the vorticity forcing by transient eddies is examined with simple linear models of the atmosphere, for instance.

The mean flow terms in the vorticity budget for the North Pacific and North Atlantic anomalous flows were not presented. Results indicate that the mean advection and stretching terms in the upper troposphere act to destroy and advect away the flow anomalies in opposition to the transient eddy terms. This is in contrast to the roles of the mean flow and anomalous eddy heat fluxes in the lower troposphere, where the eddy vorticity forcing is much smaller. Trenberth and Hurrell (1994) showed for the composited low–normal NP difference the high-frequency transient eddies act to destroy the mean temperature perturbation in a diffusive manner on an e-folding time of nearly 2 weeks. This finding is similar to the result for the North Atlantic flow anomaly as well, and it is compatible with the view that the transients are baroclinic eddies influenced by the anomalous temperature gradients on which they feed to produce downgradient transports (e.g., van Loon 1979).

As noted by HS, the real importance of the eddies is difficult to evaluate from budget computations alone. Modeling studies, such as those cited previously, perhaps allow for a better assessment of the net impact of the high-frequency transients through simulations performed with and without their presence. Nonetheless, this study points to the fact that the importance of all terms in the vorticity equation needs to be realized, especially when planetary scales of motion are considered.

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