Planetary wave–mean flow interaction in the stratosphere: a comparison between northern and southern hemispheres

By M. SHIOTANI and I. HIROTA

Geophysical Institute, Kyoto University, Kyoto 606, Japan

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SUMMARY

By the use of stratospheric height and temperature field data obtained from the TIROS satellite, dynamical interaction between planetary waves and mean zonal winds in the stratosphere is investigated. Special attention is paid to differences between northern and southern hemispheres. An analysis is made for pressure levels from 70 to 0.4 mb during the period from June 1981 to May 1982. Eliassen–Palm (E–P) flux diagnostics are used as a powerful and useful tool to investigate the wave–mean flow interaction.

From winter to summer the seasonal marches of wave activity (measured by the E–P flux) and of mean zonal wind are different between the two hemispheres. In the NH, wave activity varies intermittently with a characteristic timescale of about two weeks. Once the core of the stratospheric westerly jet shifts poleward due to a minor warming, the subsequent wave activity breaks the westerlies dramatically to replace them by the easterlies in association with the sudden warming, whilst in the SH wave activity in mid-winter is quiet, corresponding to small time variations of the maximum westerly speed. In late winter the core of the stratospheric westerly jet suddenly shifts poleward and downward due to a wavenumber-2 minor warming. After the shifting of the westerly jet the wave activity of wavenumber 1 is enhanced, and continues until early summer.

On the basis of this observational evidence we discuss: (i) the seasonal evolutions of wave activity and the mean zonal wind; (ii) the interannual variability of the stratospheric circulations; and (iii) the time variation of wave activity.

1. INTRODUCTION

It has been widely recognized that vertically propagating planetary waves excited in the troposphere are dynamically important in the stratosphere because they provide a large part of eddy momentum and heat fluxes. The propagation of planetary waves is controlled by the mean zonal wind profile, and the background flow may, in turn, be changed through the deposition of zonal momentum by planetary waves. Thus, it is important to study the properties of vertically propagating waves and the nature of their interaction with the mean zonal wind for an understanding of the general circulation of the stratosphere.

The history of wave–mean flow interaction studies began with the derivation of a ‘non-acceleration theorem’ by Eliassen and Palm (1960) for gravity waves and by Charney and Drazin (1961) for planetary waves. This theorem states that the mean zonal wind can be forced to accelerate or decelerate only when the vertically propagating waves have the effect of transience and damping. Dickinson (1969a, b) made theoretical studies on the wave–mean flow interaction problem taking account of the effects of the existence of a critical line and the presence of a Newtonian cooling process. He concluded that at least one of these effects is necessary for the forcing of zonal atmospheric motions on the planetary scale as implied by the non-acceleration theorem.

On the other hand, as a counterpart of these theoretical studies, Hirota and Sato (1969) made an analysis of day-to-day variations of planetary waves and the mean zonal wind in the NH winter stratosphere. They found that wave amplitude at 30 mb varies intermittently with a period of about two weeks and that time variations of the mean zonal wind are negatively correlated with those of the wave amplitude. Their results suggest that the transience of vertically propagating planetary waves excited in the troposphere has the effect of varying the mean zonal wind states; this wave transience is one of the conditions for violating the non-acceleration theorem.
During the 1970s the advent of observational techniques such as satellite measurements of infrared radiation enables us to investigate the stratosphere with better data coverage in both time and space (Fritz and Soules 1970; Labitzke and Barnett 1973; Barnett 1974; Hirota 1976; Hirota and Barnett 1977). Stratospheric studies based on satellite observations in that decade have been summarized by Houghton (1978), Gille (1979) and Barnett (1980).

Satellite measurements also illuminate the SH stratospheric circulation. Harwood (1975) and Hartmann (1976), paying attention to the SH stratosphere, found no major warming and found that the eastward travelling wave 2 is prominent. Moreover, Leovy and Webster (1976) made an analysis of stratospheric planetary waves in the two hemispheres, stressing the differences between the NH and the SH. To give a guide to understanding the results of this paper, we summarize briefly the notable features of the thermal field, the mean zonal winds, and planetary waves in the two hemispheres, including these results and recent observations by Hirota et al. (1983). For the thermal field:

1. Upper stratospheric temperatures in polar regions are higher in the SH than in the NH (Barnett 1974).

2. The latitudinal gradient of the zonal mean temperature at high latitudes in the SH upper stratosphere is reversed in late winter (Labitzke and Barnett 1973; Barnett 1974; Hirota et al. 1983).

For the mean zonal wind field:

1. The winter stratospheric westerlies are stronger in the SH than in the NH (Hartmann 1976; Hirota et al. 1983).

2. The stratospheric westerly jet in the SH shifts poleward and downward in late winter (Harwood 1975; Hartmann 1976). The shifting of the westerly jet is associated with the reversal of the zonal mean temperature through the thermal wind relation.

For the planetary wave:

1. Both wave 1 and wave 2 in the NH are quasi-stationary, while in the SH eastward-travelling wave 2 is dominant (Harwood 1975; Hartmann 1976) although wave 1 is quasi-stationary.

Another important stage of recent progress in theoretical studies on the wave–mean flow interaction is the ‘generalized Lagrangian-mean’ (GLM) theory of Andrews and McIntyre (1976, 1978). Motivated by their GLM theory, they reformulated the traditional wave equations into ‘transformed Eulerian-mean’ equations, which have most of the advantages of the Lagrangian-mean approach but are framed primarily in terms of observable Eulerian quantities. In this formalism a vector quantity, the so-called ‘Eliassen–Palm (E–P) flux’, is most important. For details about the transformed Eulerian-mean formalism and the motivation for the use of the E–P flux, the reader may refer to Edmon et al. (1980), Dunkerton et al. (1981) and Kanzawa (1982).

Observational studies based on the new diagnostic tool, E–P flux diagnostics, have been carried out by many authors. Among them, Palmer (1981a, b), Kanzawa (1982) and O’Neill and Youngblut (1982) obtained significant results for stratospheric sudden warmings. According to their results the sudden warmings can take place by focusing of the E–P flux into the stratospheric westerly jet located at higher latitudes in the upper or middle stratosphere. Therefore it is a necessary condition for the occurrence of sudden warmings that the stratospheric westerly jet is situated at high latitudes. Recently, Palmer and Hsu (1983) studied this preconditioning of the westerlies both on observational and theoretical bases. McIntyre and Palmer (1983, 1984) and Matsuno (1984) also studied the preconditioning on the basis of potential vorticity maps.
As regards the climatological general circulations in the stratosphere, Smith (1983) and Geller et al. (1983) applied the E–P flux diagnostics to the monthly mean field in the NH winter. Hartmann et al. (1984) also investigated observational features of the wave–mean flow interaction in the SH using the E–P flux diagnostics.

However, these studies were carried out for limited periods (such as sudden warmings) or for a limited region (one of the two hemispheres). Making good use of satellite measurements we should grasp global pictures of the stratospheric circulation and the seasonal evolutions of wave activity and the mean zonal wind. In this sense, it is stressed that comparisons of dynamical features between the NH and the SH provide a valuable insight into the nature of stratospheric circulations; in other words, two hemispheres provide a kind of controlled experiment. Moreover, the seasonal march of wave activity and the mean zonal wind must be one of the most interesting problems in the dynamics of the stratosphere. For example, a sudden warming in the NH, in the sense that the westerly jet is broken to easterlies, has never been reported in early winter (November and December); we should understand the sudden warming event as it is realized in the seasonal march of wave activity and the mean zonal wind.

With this in view, Hirota et al. (1983) recently studied the difference of upper stratospheric circulations in the NH and the SH, and confirmed some evidence as listed earlier. In addition, they found that the seasonal evolutions of wave activity in the two hemispheres are different: from winter to spring wave activity is more vigorous in the NH than in the SH; whereas from spring to summer, more vigorous in the SH than in the NH. In this paper, we will present some observational evidence of wave–mean flow interaction in the stratosphere by expanding this work.

2. DATA AND METHOD OF ANALYSIS

In this study we use global stratospheric data supplied by National Meteorological Center (NMC). This data set consists of geopotential heights and temperatures for the constant pressure levels 70, 50, 30, 10, 5, 2, 1 and 0.4 mb, covering the geometric height range 18 to 55 km, approximately. Analysis was made for the year June 1981 to May 1982.

The NMC data are originally from the TIROS operational vertical sounder (TOVS) system which derives stratospheric soundings from nine stratospheric channels on three instruments: stratospheric sounding unit (SSU), high resolution infrared sounder (HIRS), and the microwave sounding unit (MSU). This system provides layer-mean temperatures between standard pressure levels; geopotential heights are derived through the hypersometric equation by adding layer-mean thicknesses to a lower boundary height at 100mb; (for details, see Geller et al. 1983).

We converted the original NMC data, which are on polar stereographic grids, onto latitude–longitude grids (5°×5°) and made a harmonic analysis along each latitude circle. Physical quantities due to eddies, such as E–P flux and its divergence, were calculated for zonal wavenumbers 1 to 6.

Following Kanzawa (1982) (notations are based on Holton 1975 and Dunkerton et al. 1981), we define the E–P flux $F$ for the quasi-geostrophic motions in spherical coordinates $(\lambda, \theta, z = -H \ln (p/p_s))$ by

$$F = (F(\theta), F(z))$$

(1)

where

$$F(\theta) = -\rho_0(z) a \cos \theta \overline{u'v'}$$

(2)

$$F(z) = +\rho_0(z) a \cos \theta \overline{(f/N^2)v'\Phi_z}$$

(3)
\( \rho_0(z) = \rho_0 = \rho_0 \exp(-z/H) \). We also define 'residual meridional circulation' \((\vec{v}^*, \vec{w}^*)\) by
\[
\vec{v}^* = \vec{v} - \frac{1}{\rho_0} \frac{\partial}{\partial z} \left( \rho_0 \vec{v} \frac{\Phi_z}{N^2} \right)
\]
\[
\vec{w}^* = \vec{w} + \frac{1}{a \cos \theta} \frac{\partial}{\partial \theta} \left( \cos \theta \frac{\vec{v} \frac{\Phi_z}{N^2}}{a^2 \vec{w}} \right).
\]

Then we can transform traditional equations into:
\[
\frac{\partial \vec{u}}{\partial t} - f\vec{v}^* = \left( \frac{1}{\rho_0 a \cos \theta} \right) \nabla \cdot \vec{F}
\]
\[
\frac{\partial \Phi_z}{\partial t} + N^2 \vec{w}^* = 0
\]
\[
(1/a \cos \theta) \frac{\partial}{\partial \theta} \left( \vec{v}^* \cos \theta \right)/\partial \theta + (1/\rho_0) \frac{\partial \rho_0 \vec{w}^*}{\partial z} = 0
\]

where
\[
\nabla \cdot \vec{F} = \left( \frac{1}{a \cos \theta} \right) \frac{\partial (F(\theta) \cos \theta)}{\partial \theta} + \frac{\partial F(z)}{\partial z}.
\]

The vector quantity \( \vec{F} \) and its divergence \( \nabla \cdot \vec{F} \) play a central role in this formalism. From the theoretical consideration as in Kanzaw (1982) we can interpret \( \vec{F} \) as westward angular wave momentum flux propagating through westerly flow and \( \nabla \cdot \vec{F} \) as wave-induced torque per unit volume acting on the mean flow. In the transformed formalism the non-acceleration theorem is simply represented as \( \nabla \cdot \vec{F} = 0 \). To calculate the E–P flux we assume \( H \) and \( N \) to be constant \( (H = 7 \times 10^3 \text{m}, N = 2 \times 10^{-2} \text{s}^{-1}) \), which correspond to the isothermal atmosphere with \( T_i = 239 \text{K}. \)

For representing the E–P flux in vertical cross-sections we followed the graphical conventions as described by Edmon et al. (1980). According to their remarks we should take account of the spherical geometry of earth's surface. We define the coordinate \( y = a \theta \) and form the product of \( \vec{F} \) and the area factor \( \cos \theta \):
\[
\vec{F}^* = \vec{F} \cos \theta
\]
and also define its divergence as
\[
\nabla \cdot \vec{F}^* = (\nabla \cdot \vec{F}) \cos \theta
\]
where
\[
\nabla \cdot \vec{F}^* = \frac{\partial F^*(y)}{\partial y} + \frac{\partial F^*(z)}{\partial z}.
\]

Because the scales of the coordinates \( y \) and \( z \) are different, we multiply by a constant \( c \) corresponding to \( \partial y/\partial z \) (in this study, \( c = 125 \)). If we define \( \vec{E}^* \) by
\[
\vec{E}^* = (F^*(y), cF^*(z))
\]
then \( \vec{E}^* \) appears to be nondivergent if and only if \( \nabla \cdot \vec{F} = 0 \).

In addition, to see the effect of waves on the time change of the mean zonal wind, we define \( D_F \) as
\[
D_F = (1/\rho_0 a \cos \theta) \nabla \cdot \vec{F}.
\]
Then we can rewrite Eq. (6) as
\[
\frac{\partial \vec{u}}{\partial t} - f\vec{v}^* = D_F.
\]

This shows that \( D_F \) can be regarded as a zonal force per unit mass acting on the mean state. In the following cross-sections we use the flux \( \vec{E}^* \) and the zonal force per unit mass \( D_F \) instead of the original flux \( \vec{F} \) and its divergence \( \nabla \cdot \vec{F} \).
As a measure of the wave activity we define $|\mathbf{F}|$, from the theoretical consideration as in Palmer (1982), as

$$|\mathbf{F}| = [\{F(\theta)\}^2 + \{(N/f)F(z)\}^2]^\frac{1}{2}. \quad (16)$$

A deformation factor $N/f$ appears because we should treat the two components of the E–P flux isotropically. The deformation factor is evaluated as 158 for $\theta = 60^\circ$; this is nearly equal to the scaling factor $c (= 125)$. Therefore we can regard the E–P flux represented in this study as having been treated approximately isotropically. In addition, from Eqs. (2) and (3) we see that $|\mathbf{F}|$ is proportional to the wave amplitude squared.

3. Results

(a) Seasonal march

First we describe the seasonal march of the mean zonal wind and that of the E–P flux. For the purpose of studying the wave–mean flow interaction we make an analysis for the period of vigorous wave activity; in practice we select six months — December 1981 to May 1982 for the NH and June 1981 to November 1981 for the SH.

One of the most important results of Hirota et al. (1983) is that the seasonal march of zonal mean temperatures and wave amplitudes is different between the NH and the SH in the upper stratosphere. They showed this by simple statistics of standard deviations with respect to time for two seasons: ‘winter to spring’ and ‘spring to summer’. In correspondence with their result the period of the first three months (for the NH, December to February and for the SH, June to August) is called ‘winter to spring’ and the period of the second three months (for the NH, March to May and for the SH, September to November) is called ‘spring to summer’.

(i) Mean zonal wind. Figure 1 shows latitude–time sections of the mean zonal wind at the 1 mb level, estimated geostrophically. We selected this level to see the poleward shifting of the maximum westerlies clearly. From Fig. 1 we can see the following features: For the NH (Fig. 1(a)):

1. From December to early January the maximum westerlies, with a speed of about 100 m s$^{-1}$, are slowly shifting poleward.
2. Around 8 January a minor warming occurs and after that the maximum westerlies are established at high latitudes, about 60$^\circ$N (16 January).
3. Then a major warming occurs and the westerlies are reversed to easterlies at high latitudes (around 26 January).
4. In February the westerlies are re-established but the jet profile is not so clearly formed as seen in December. Summer easterlies appear at this level in April to the north of 60$^\circ$N and in May to the south of 60$^\circ$N.

For the SH (Fig. 1(b)):

1. From June to July the maximum westerlies, with a speed of about 130 m s$^{-1}$, stronger than in the NH, are sharply located at around 40$^\circ$S and their time variations are small.
2. In mid-August the maximum westerlies shift poleward about 20$^\circ$ in about a week.
3. From the shifting (around August 15) to October the maximum westerlies are located at around 60$^\circ$S and their time variations in this period are larger than those in the pre-shifting period.
4. In early November summer circulations are established for all latitudes simultaneously.
(ii) $E-P$ flux. Next we describe the seasonal march of the $E-P$ flux for the two hemispheres. Cross-section analyses will be shown later. In this section we regard the $E-P$ flux as a measure of the net propagation of wave activity. Figure 2 shows latitude-time sections of the $E-P$ flux at the 5 mb level for total wavenumber (1–6) contributions. We selected this level to see the switching of the $E-P$ vectors clearly. From Fig. 2 we can see the following features:

For the NH (Fig. 2(a)):

1. From December to March wave activity varies intermittently with a rhythm of about two weeks; such a quasi-periodic change has been reported earlier by Hirota and Sato (1969) and Madden (1975) from the analysis of wave amplitude variations.

2. For most of the period in vigorous wave activity the $E-P$ vectors point equatorward and upward, with their main sources around 60$^\circ$N. These features agree well with previous results for the normal winter stratosphere using $E-P$ flux diagnostics (e.g. Smith 1983; Geller et al. 1983).

3. However, when the major warming occurs in late January, the $E-P$ vectors point strongly upward with weak poleward components. Notice that there is little difference in the magnitude of the wave activity from one event to another. Thus, it is not true that the most vigorous wave activity brings about a major warming; more important is the direction of the $E-P$ flux.

For the SH (Fig. 2(b)):

1. From June to July wave activity is relatively weak, corresponding to the strong and sharp westerly jet at the same period. As regards the wave–mean flow interaction it is interesting to note the relation between the weak wave activity and the small time variations of the maximum westerlies.
Figure 2. Latitude–time sections of the total E–P flux at the 5 mb level from winter to summer for (a) the NH and (b) the SH. The arrow scale and scaling factor $c$ (see section 3) are presented in the figure.
Figure 3. Latitude–time sections of the E–P flux for wavenumbers 1 (lower) and 2 (upper) at the 5 mb level for selected three-month periods: (a) from December 1981 to February 1982 in the NH and (b) from August 1981 to October 1981 in the SH. The arrow scale and scaling factor c (see section 3) are presented in the figure.

(2) In August wave activity increases, and varies intermittently with a rhythm of about 10 days. From September to October wave activity goes on continuously, not intermittently as in the NH. The magnitude of the wave activity is rather less in the SH than in the NH.
(3) For most of the period in vigorous wave activity the E–P vectors point equatorward and upward as in the NH. Also in the SH, there are occasions when the E–P vectors point strongly upward with weak poleward components (around 17 September). However, unlike the NH, a major warming event in which westerlies are replaced by easterlies does not occur.

To see the role of each wavenumber, we present a picture similar to Fig. 2 but for wavenumbers 1 and 2 separately, for a selected period of three months, for the NH December to February (Fig. 3(a)), and for the SH August to October (Fig. 3(b)). Referring to Fig. 2 we see from Fig. 3 that almost all the contribution is from wavenumbers 1 and 2 components; and that the periods of vigorous wave activity for waves 1 and 2 do not overlap in general.

In the NH (Fig. 3(a)) if we consider the E–P flux separately for each wavenumber contribution, the rhythm of the wave-1 activity is interrupted in late January; however, at around the same time wave-2 activity rises and compensates for the falling wave-1 activity. Thus in the total wavenumber contribution we can see the rhythm of wave activity with a timescale of about 2 weeks. It seems that the wave-1 and wave-2 activities in the NH appear with no obvious pattern in time.

On the other hand, in the SH, the times of wave activity for waves 1 and 2 are different (Fig. 3(b)); wave-2 activity lies mainly in August, whereas wave-1 activity lies from late August (just after the shifting of the westerly jet) to mid-October. As was shown in earlier studies, planetary wave properties for both hemispheres are different; in the NH waves 1 and 2 are quasi-stationary, while in the SH eastward travelling wave 2 is prominent though wave 1 is quasi-stationary. Regarding the different wave properties of waves 1 and 2 in the SH, it is interesting to see that wave-1 activity begins in association with the poleward shifting of the westerly jet.

(b) Vertical cross-section analysis

In section 3(a) we observed how the mean zonal wind and the E–P flux vary in the seasonal march at one level. Next we make a vertical cross-section analysis to see the behaviour of the mean zonal wind and the E–P flux in the whole stratosphere. Our main interest is in comparison of the two hemispheres. Figures 4 to 8 show sequences of latitude–height cross-sections of the total E–P flux and its divergence expressed as the zonal force per unit mass ($D_F$) (left), and the mean zonal wind (right). Figure 4 is presented every five days and Figs. 5 to 8 are presented every three days. (We have made a 6-minute, 16 mm colour movie which shows the seasonal evolution of the mean zonal wind and the E–P flux; this is available at cost from the authors.)

(i) The rhythm of minor warmings in the NH. Figure 4 shows an example of a sequence for a minor warming in the NH. On 28 November wave activity is weak while the mean zonal winds are strong in high latitudes, especially in the lower stratosphere. Then wave activity is enhanced and the mean zonal winds in high latitudes are weakened, resulting in the appearance of easterlies there (3 December); however, the speed of the stratospheric westerly jet is little changed but contour lines in the poleward flank of the westerly jet are packed. In terms of zonal mean potential vorticity ($\vec{\eta}$) (e.g. recent studies by McIntyre (1982) and Matsumo (1984)), we can interpret these features in the mean zonal wind change as a decrease of $\vec{q}$ in high latitudes and an increase in middle latitudes; that is, propagating planetary waves bring a change in the distribution of $\vec{q}$.

Around 3 December, the E–P vectors point strongly upward in the lower stratosphere, bend equatorward in the middle stratosphere and point nearly equatorward in the upper stratosphere. Values of $D_F$, representing the effect of waves on the mean flow,
Figure 4. Latitude–height sections of the total E–P flux and its divergence expressed as the zonal force per unit mass, $D_f$, (left) and the mean zonal wind (right) for the NH from 28 November to 13 December 1981. The arrow scale and scaling factor $c$ are presented in the figure. The contour interval for $D_f$ is $10^{-4}$ m s$^{-2}$; negative values are shaded. Absolute values of the wind speed are shown by tone patterns (see the tone bar, units: m s$^{-1}$). Easterlies are over oblique lines.
Figure 5. As Fig. 4 but for the NH from 3 to 12 January 1982.

are negative in almost the whole stratosphere. The pattern of the E–P vectors agrees well with all previous results for the normal winter in the NH stratosphere (e.g. Smith 1983; Geller et al. 1983).
From around 8 December wave activity is gradually weakened, then the westerlies in high latitudes are strengthened again; contour lines in the poleward flank of the westerly jet are gradually spread. On 13 December when there are no strong upward E–P vectors in the lower stratosphere the E–P vectors in the middle stratosphere point nearly equatorward. In addition, the distribution of \( D_F \) shows a dipole pattern, i.e. positive at high latitudes and negative at middle or lower latitudes. Again, in terms of \( \bar{q} \), we can interpret this returning back process of the mean zonal wind as a redistribution of \( \bar{q} \), i.e. an increase of \( \bar{q} \) in high latitudes and a decrease in middle latitudes. Because \( \nabla \cdot \mathbf{F} \) is related to the meridional flux of potential vorticity as shown in Edmon et al. (1980) and Kanzawa (1982), this dipole pattern is consistent with time changes of \( \bar{q} \).

As an interpretation of the near-equatorward E–P flux and the dipole pattern of \( D_F \), Hartmann et al. (1984) suggested that, on the basis of their observations in the SH, these configurations are due to an in situ source of wave activity, such as large-scale shear instability, whereas Palmer and Hsu (1983) stated that the nonlinear interaction between waves 1 and 2 could give rise to such a dipole pattern, in a flow that was internally stable.

Positive \( D_F \) works to strengthen the mean zonal wind in the high latitude stratosphere (13 December), then wave activity becomes large. In this way a minor warming in the NH repeats itself.

(ii) Shifting of the westerly jet. As seen in Fig. 1 the maximum westerlies in both hemispheres tend to shift poleward in their seasonal evolutions. The shifting of the stratospheric westerly jet is important for understanding stratospheric circulations in the two hemispheres. Hence we describe the behaviour of the mean zonal wind and the feature of the E–P flux around the period of the shifting in the two hemispheres.

In the NH (Fig. 5), the stratospheric westerly jet in the mid-latitude upper stratosphere is weakened due to a minor warming (after January 3). We see large \( D_F \) regions in the mid-latitude stratosphere on 3 January and in the high latitude stratosphere on 6 January. On 9 January, although the upward E–P flux exists in the lower stratosphere, the E–P flux in the middle stratosphere points nearly equatorward and the distribution of \( D_F \) shows a dipole pattern, as seen in Fig. 4 (on 13 December). This means that the westerlies are accelerated at high latitudes and decelerated at mid-latitudes. Then, on 12 January the stratospheric westerlies are established at high latitudes. We can interpret these features as a redistributing process of potential vorticity.

In the SH the poleward and downward shifting of the stratospheric westerly jet in late winter has been reported by many authors, e.g. Leovy and Webster (1976), Hartmann (1976) and Hartmann et al. (1984) for the 1971 winter, the 1973 winter and the 1979 winter, respectively. However, they paid little attention to day-to-day variations of the mean zonal wind and wave activity.

From Fig. 6 we can see the poleward and downward shifting of the westerly jet; it takes about 10 days. Relating to the wave activity we can conclude that a minor warming (12 to 15 August) which leads to the negative \( D_F \) at mid-latitudes and the positive \( D_F \) at high latitudes contributes to the shifting of the westerly jet. The scale of the wave activity and values of \( D_F \), which leads to the shifting of the westerly jet are less in the SH than in the NH. Time variations of the maximum westerlies are also smaller in the SH than in the NH. Again, we are interested in the distributions of the E–P flux and \( D_F \) on 15 August, a so-called dipole pattern, also seen in the NH.

A complementary (but compatible) idea of the physical cause of the shifting of the westerly jet is presented by McIntyre (1982) and McIntyre and Palmer (1983, 1984), in terms of erosion of the potential vorticity distribution in the polar vortex.

(iii) Episodes after the shifting of the westerly jet. With regard to the shifting of the
Figure 6. As Fig. 4 but for the SH from 9 to 18 August 1982.

westerly jet the seasonal march of the mean zonal wind in the two hemispheres is similar; however, episodes after the shifting are quite different.
1.0 \times 10^7 \text{kg s}^{-2} : c = 125

Figure 7. As Fig. 4 but for the NH from 18 to 27 January 1982.
Figure 8. As Fig. 4 but for the SH from 14 to 23 September 1982.
In the NH (Fig. 7) the stratospheric westerly jet situated in the high latitude upper stratosphere (18 January) leads to the focusing of the E–P flux (18 to 24 January). Before the shifting of the westerly jet the E–P vectors point upward and equatorward, but after the shifting they point strongly upward and, at high latitudes, poleward. Negative $D_F$ regions are observed for almost the whole stratosphere and maximum values are $9 \times 10^{-4} \text{ m s}^{-2}$ in the high latitude stratosphere on 24 January. Then the westerlies are dramatically broken to the easterlies; this is a so-called major warming.

In the SH, on the other hand, after the shifting the stratospheric westerly jet continues to be stably situated in the high latitude middle stratosphere for about two months. The E–P flux occasionally focuses into high latitudes, but the westerlies are not broken to the easterlies. Figure 8 is just one example in 1981 to show the focusing of the E–P flux into high latitudes (17 September). The maximum values of $D_F$ are about $6 \times 10^{-4} \text{ m s}^{-2}$ in the high latitude stratosphere, weaker than the $9 \times 10^{-4} \text{ m s}^{-2}$ in the NH on 24 January. However, no major warming occurs and no change of its position.

(c) Wave–mean flow interaction

In this sub-section, we will discuss the relationship between the E–P flux, its divergence expressed as the zonal force per unit mass, $D_F$, and the time changes of the mean zonal wind. Figure 9 shows an example of latitude–time sections of the E–P flux, $D_F$, and $\partial \bar{u} / \partial t$ at the 5 mb level for three selected months in the NH. To see the gross features of $D_F$ and $\partial \bar{u} / \partial t$, we applied a 5-day running mean to their time series.

First, we notice the relation between $D_F$ and $\partial \bar{u} / \partial t$. The zonal force per unit mass ($D_F$) implies the effect of the waves on the mean flow; moreover, from the theoretical prediction by Matsumo and Nakamura (1979) and Dunkerton et al. (1981), poleward residual flow $\bar{v}^*$ arises when the mean flow is decelerated. Therefore the sense of variation in $D_F$ and $\partial \bar{u} / \partial t$ is expected to be the same near a maximum of $D_F$. Indeed, Figs. 9(b) and (c) show that this is the case in middle and high latitudes. To confirm this relation we calculate the correlation coefficients between the time series of $D_F$ and $\partial \bar{u} / \partial t$ ($\text{COR}(D_F, \partial \bar{u} / \partial t)$) at each latitude and pressure level.

Figure 10 shows latitude–height sections of the coefficient $\text{COR}(D_F, \partial \bar{u} / \partial t)$ for six months from winter to summer in the two hemispheres. In the NH (Fig. 10(a)) the coefficient is positive and significantly large at high latitudes, with the maximum value in excess of 0.8 in the upper stratosphere, and it decreases toward the equator. In the SH (Fig. 10(b)) the coefficient is also large at around 60°S and 40°S in the upper stratosphere with maximum values in excess of 0.5, though it is smaller than in the NH. The larger value of the coefficient in the NH is perhaps due to the larger time change of the mean zonal wind in the NH than in the SH. The two maxima in the SH reflect the position of the westerly jet before and after the shifting. The result for the SH is in good agreement with similar statistics by Hartmann et al. (1984).

Next we examine the relationship between the two terms $|F|$ and $D_F$ (Figs. 9(a) and (b)). It is clear that at the 5 mb level the term $D_F$ has a large negative value when the wave activity is vigorous. That is, the two terms $|F|$ and $D_F$ are in good negative correlation. On the observational basis Hirota and Sato (1969) showed that wave amplitudes and mean zonal wind are in good negative correlation. Their results are commonly accepted as a good representation of the effect of wave transience; however, based on their results $|F|$ (regarded as a measure of wave amplitude) should be in negative correlation with the mean zonal wind $\bar{u}$, namely from Fig. 10, $-\partial D_F / \partial t$.

To clarify the relation between wave activity and the mean zonal wind we calculate correlation coefficients between the time series of $|F|$ and $\bar{u}$ ($\text{COR}(|F|, \bar{u})$) and those between the time series of $|F|$ and $\partial \bar{u} / \partial t$ ($\text{COR}(|F|, \partial \bar{u} / \partial t)$) at each latitude and pressure
Figure 9. Latitude–time sections of (a) the total E–P flux; (b) its divergence expressed as the zonal force per unit mass \( \langle D \rangle \); and (c) time changes of the mean zonal wind \( \langle \delta \bar{u} / \partial t \rangle \) at the 5 mb level from December 1981 to February 1982 in the NH. The contour intervals of \( \langle D \rangle \) and \( \langle \delta \bar{u} / \partial t \rangle \) are 4.0 m s\(^{-1}\) day\(^{-1}\) and 2.0 m s\(^{-1}\) day\(^{-1}\) respectively. Negative values are shaded.

level for six months from winter to summer. Figure 11 shows latitude–height sections of the coefficient \( \text{COR}(|F|, \delta \bar{u} / \partial t) \) and Fig. 12 shows those of the coefficient \( \text{COR}(|F|, \bar{u}) \). To calculate the coefficient \( \text{COR}(|F|, \bar{u}) \) we remove long-term variations of the mean zonal wind \( \bar{u} \) by a high pass filter. The numerical filter was designed to have a half power period of 39 days.

Figure 11 shows that the most negative values, less than \(-0.4\), are observed in the middle latitude upper stratosphere for both hemispheres; this is confirmation of the relationship between the terms \( |F| \) and \( D_f \) (Figs. 9(a) and (b)). On the other hand, in Fig. 12(a) for the NH we can see two minima, less than \(-0.4\) in the high latitude lower stratosphere and less than \(-0.5\) at the middle latitude stratopause level. The negative correlation in the high latitude lower stratosphere supports the result of Hirota and Sato (1969). In Fig. 12(b) for the SH we can see the minimum, less than \(-0.3\) in the mid-latitude middle stratosphere.

It seems that the coefficients in Figs. 11 and 12 are rather small. However, this is because we selected a long period (180 days) from winter to summer for the statistics. If we select a limited period in which the wave–mean flow interaction typically takes place, these coefficients become more confident. For example, in the statistics of \( \text{COR}(|F|, \delta \bar{u} / \partial t) \) for 90 days from December to February for the NH and from August to October for the SH, the most negative values become \(-0.67\) and \(-0.68\), respectively.
Figure 10. Latitude–height sections of correlation coefficients between $D_F$ and $\partial \bar{u}/\partial t$, $\text{COR}(D_F, \partial \bar{u}/\partial t)$, for (a) the NH and (b) the SH.

We suppose that the negative correlation between the two terms $|\mathbf{F}|$ and $\bar{u}$ in the high latitude lower stratosphere of the NH is probably due to the transience of vertically propagating planetary waves. Figures 11 and 12 suggest, however, that this is not always the case in the whole stratosphere. It is interesting to note that the configurations of $\text{COR}(|\mathbf{F}|, \partial \bar{u}/\partial t)$ in the two hemispheres are similar (Fig. 11); the most negative values are observed in the middle latitude upper stratosphere. In connection with recent studies of wave breaking (McIntyre and Palmer 1983, 1984), the deceleration of the mean zonal wind in these regions may be associated with the dissipation due to the wave transience of small scales. It is a further problem to study a physical mechanism for the wave–mean flow interaction in the middle latitude upper stratosphere. The negative correlation between $|\mathbf{F}|$ and $\bar{u}$ at the mid-latitude stratopause level in the NH is also still open to question.

4. CONCLUDING REMARKS

(a) Schematic pictures

The results presented in section 3 are based on only one year’s statistics; however, most of the characteristics of the mean zonal wind and the wave activity for the NH and the SH are generally supported by many observations for other years.
Figure 11. As Fig. 10 but for the coefficient COR(|F|, \partial u/\partial t) in (a) the NH and (b) the SH.

Figure 12. As Fig. 10 but for the coefficient COR(|F|, \bar{u}) in (a) the NH and (b) the SH.
For example, the rhythm of time variations in the mean zonal wind and the wave activity in the NH, with a typical time scale of two weeks, has been reported by Hirota and Sato (1969) and Madden (1975). The result in this paper gives a modern version of the interpretation of the relation between the mean zonal wind and wave activity using a new diagnostic tool: the E–P flux.

The poleward shifting of the westerly jet in the NH, which is important for preconditioning a sudden warming, has been reported by Quiroz et al. (1975), Kanzawa (1980, 1982) and Palmer (1981a, b). For the SH the poleward and downward shifting of the westerly jet, which is one of the most important aspects in the SH, has been reported by Hartmann (1976) and Leovy and Webster (1976). Hirota et al. (1983) have shown the shifting in terms of the reversal of the meridional temperature gradient in the upper stratosphere for four years.

An enhancement of the wave-1 activity after the shifting of the westerly jet in the SH has been first pointed out clearly in this paper. A similar feature has been reported recently by Hartmann et al. (1984) for the 1979 winter. It is also interesting to see that the period of weak wave activity in the SH mid-winter is simulated in a simplified general circulation model by Holton and Wehrbein (1980).

Concerning the question of interannual variability, Holton and Tan (1982) and Labitzke (1982) have discussed the relationship between the extratropical circulation in the NH stratosphere and the equatorial zonal wind change, i.e. the quasi-biennial oscillation. In the SH there is no major warming; however, it is interesting to see the relation between the time of the shifting of the westerly jet and the wind regime of the quasi-biennial oscillation (Table 1). We will give an interpretation of this table in subsection (b) below.

Apart from the interannual variability in the two hemispheres, our present results for one year, 1981–1982, give some characteristics of the seasonal march of the mean zonal wind and wave activity, as shown in a schematic picture (Fig. 13). The features in Fig. 13 are representative of the mid-latitude upper stratosphere.

### Table 1: Relation between the Time of the Shifting of the Stratospheric Westerly Jet in the SH and the Wind Regime of the Quasi-Biennial Oscillation at the 50mb Level

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The wind regime of the QBO (second) is derived from Fig. 3 in Labitzke (1982). Notations are; W: westerly; E: easterly; : weak zonal wind; —: undetermined. The time of the shifting is depicted by the symbol *. It is classified as the first, middle and last part of a month. For the first four observations the time of the shifting is deduced from the thermal field in the upper stratosphere. For reference, data source and observers are listed; lack of observer means that the data source belongs to our laboratory (Kyoto).
For the NH (upper):
(1) Wave activity, with contributions from both wave 1 and wave 2, varies erratically from winter to spring, with no systematic pattern but with a characteristic time scale of about two weeks.
(2) The time change of the mean zonal wind \( \partial \bar{u} / \partial t \) is in negative correlation with the wave activity.
(3) In mid-winter the stratospheric westerly jet shifts poleward owing to the occurrence of a minor warming. After the shifting, subsequent wave activity brings about a major warming.

For the SH (lower):
(1) In winter time wave activity is weak, resulting in the small time changes of the mean zonal wind.
(2) In late winter the stratospheric westerly jet shifts poleward and downward in association with the wave activity.
(3) After the shifting, the wave-1 activity is enhanced and continues till late spring. In this period the time change of the mean zonal wind \( \partial \bar{u} / \partial t \) is, as in the NH, in negative correlation with the wave activity.

Next we describe schematically how the mean zonal wind profile changes in time evolutions. Figure 14 shows sequences of wind profile changes in vertical cross-sections. For the NH (Fig. 14(a)):
(1) The stratospheric westerlies in high latitudes change their speeds with a characteristic time scale of two weeks (A ↔ B); that is, the rhythm of minor warmings.
(2) At some time the westerly jet is established in the high latitude stratosphere after a minor warming (B→C).
(3) Then the propagating wave through this wind profile breaks down the westerlies to easterlies (C→D); this is a so-called major warming. After a major warming the wind profile does not necessarily return back to state C.

For the SH (Fig. 14(b)):
(1) The strong and sharpened westerly jet exists in mid-winter (A).
(2) In late winter the westerly jet shifts poleward and downward in association with the wave activity (A→B).
(3) After the shifting, unlike the NH, the westerly jet does not change its position (in the high latitude middle stratosphere) but oscillates in speed (B→C).

(b) Interannual variability

Concerning the interannual variability the relation between the quasi-biennial oscillation in the tropical region and the seasonal march of the mean zonal wind in the stratosphere is suggestive in both hemispheres. In particular, this problem is interesting in relation to the shifting of the westerly jet. The shifting is essential for the two hemispheres in the stratospheric general circulations, although the evolution of the mean zonal wind after the shifting is different between the hemispheres.

Table 1 shows the relation between the time of the shifting of the westerly jet in the SH and the wind regime of the quasi-biennial oscillation. From this table it appears that the equatorial wind regime and the time of the shifting in the SH are closely connected. In general, when the equatorial zonal winds at 50 mb are easterly the time of the shifting
is early (July) and when they are westerly it is late (August or September). On the other hand, Labitzke (1982) noted that mid-winter major warmings in the NH occur readily when the equatorial zonal wind regime is easterly. These features are interpreted as follows: when the regime of the quasi-biennial oscillation is easterly the poleward shifting is earlier for both hemispheres, and particularly in the NH it leads to the mid-winter stratospheric sudden warming.

(c) Troposphere–stratosphere coupling

Finally, we discuss possible mechanisms to explain the time variation of wave activity. In our results, there are two clues to explain the mechanism: (i) in the NH, in the rhythm of minor warmings wave activity rises when the westerlies in the high latitude lower stratosphere are enhanced; (ii) in the SH, after the shifting of the westerly jet, which provides an enhancement of the westerlies in the high latitude lower stratosphere, wave activity of wave 1 rises and continues until the westerlies are weakened. In view of these results, we can say that wave momentum flux is strongly connected with the wind profile in the lower stratosphere and perhaps also in the troposphere. Thus we present two hypotheses concerning the mechanism of variations of wave activity:

1. A direct excitement of wave activity by the enhanced westerlies in the high latitude troposphere; e.g. the excitation of tropospheric wave activity by the orographic effect.
2. A ‘shutter’ mechanism at around tropopause level, which assumes constant wave activity in the troposphere. When the westerlies are enhanced at the high latitude tropopause level the shutter is open and tropospheric waves easily penetrate into the stratosphere. Then, because of wave–mean flow interaction the westerlies are weakened and the shutter is closed to prohibit tropospheric activity entering the stratosphere.

Figure 15 illustrates the two hypotheses, stressing the region of the lower stratosphere and the troposphere. In (a) (the first hypothesis), the stronger the westerlies in the high latitude troposphere are, the more tropospheric wave activity is excited; the wave is free to propagate into the stratosphere.

In (b) (the second hypothesis) even if wave activity at the bottom of the troposphere is constant the direction of wave propagation is controlled by the wind profile around the tropopause level. When the westerlies around the high latitude tropopause level are weak the E–P vectors point to the tropospheric westerly jet; when the westerlies at

![Schematic vertical cross-sections of the mean zonal wind and the wave activity for two hypotheses](see text).
high latitudes are enhanced the E–P vectors are guided to the stratospheric maximum westerlies. This mechanism is somewhat similar to that of stratospheric vacillation models (Holton and Mass 1976; Holton and Dunkerton 1978; Schoeberl 1983). Recently, O’Neill and Youngblut (1982) tried to explain the amplification of wave activity which causes sudden warmings using a mechanism similar to the second hypothesis.

For both hypotheses the mechanism for weakening westerlies is considered to be interaction between the mean zonal wind and propagating waves. However, the mechanism to accelerate the westerlies, as observed in the seasonal evolution of the mean zonal wind or in the situation such as the dipole pattern of the E–P flux divergence, is not yet clear. To have further insight into the stratospheric dynamics we should make more extensive analyses of the wind field and the planetary waves, including the troposphere.

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