A Transformed Eulerian Mean Diagnostics for a Sun-Weather Relation

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Abstract

The aim of this study is to find a linkage of atmospheric responses to the solar sector boundary (SSB) passages (Wilcox *et al.*, 1974; Misumi, 1981; 1983). Variations of both zonal mean fields and planetary waves around the SSB passages are investigated using the transformed Eulerian mean (TEM) equations system including the refractive index. Deviations from the time mean field in 8 winters from 1964 to 1972 are mainly discussed.

It is found that the Eliassen-Palm (EP) flux for zonal wavenumber 2 and 3 in the troposphere begins to decrease about 5 days prior to SSB passages. Such decrease leads to the divergence of the EP flux near the tropopause in high latitudes. This is responsible for the temperature decrease in the polar troposphere, and consequently for the decrease of the vorticity area index in the upper troposphere. The decreases of the EP flux for both wavenumber 2 originating from the troposphere and wavenumber 1 being at the 150 mb level on 4 days before the SSB passages seem to create the minimum of the polar temperature in the middle stratosphere.

Among above mentioned processes, only the decrease of the EP flux is not diagnosed by the TEM equations system. The decrease of the EP flux in the troposphere is statistically significant and is seen only in the same period that other atmospheric responses occur. This type response would be due to the external modulation.

1. Introduction

The study of atmospheric dynamics respondent to the solar sector boundaries (SSB) passage was carried out by Wilcox *et al.* (1974; hereafter referred to as W74). They pointed out that the vorticity area index (VAI) defined by Roberts and Olson (1973) had a minimum on the first day after each SSB passage in the winters of 1964–1970. Wilcox *et al.* (1976) obtained the same result for the extended winters of 1963–1974. Hines and Halevy (1977) supported this kind of response from a statistical point of view.

Pitcock and Shapiro (1982) reexamined those results and questioned about the reality of such response to the SSB passage. The es-

sence of their criticism consists of two points; (1) The VAI is an artificial parameter. No response of physical parameters such as the kinetic energy and absolute vorticity was found (Bhatnager and Jakobsson, 1978; Williams, 1978; Williams and Gerety, 1980). (2) The reported VAI response itself can not be found before 1963 or after 1973. (Williams and Gerety, 1978; Shapiro, 1979; Arora and Padgaonkar, 1981)

Recently, Misumi (1983; hereafter referred to as M83) analyzed zonal mean temperature and height on 7 pressure levels below the 100 mb level in the Northern Hemisphere and found that the temperature in the polar troposphere reached a minimum on each SSB passage. Though this occurred in a limited area, the magnitude of the response was extremely significant in 8 winters (Dec., Jan., and Feb.)

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from 1964 to 1972. Misumi (1981; hereafter referred to as M81) also reported minimum temperature in the polar middle stratosphere on 2.5 day after the SSB passages in the same winters. Since temperature is a fundamental parameter, these results answer Pittock and Shapiro's criticism (1). It should be mentioned that on VAI response to the geomagnetic storm is seen for longer period than that to the SSB passages (Arora and Padgaonkar, 1981). However, that response to geomagnetic storm is not so significant in a statistical view. Therefore, we should investigate more significant sun-weather relation.

M83 suggested that the dominant cause of the temperature response might be a dynamical process attributed mainly to the planetary wave. The transformed Eulerian mean formulation was introduced in order to describe wave-mean flow interactions by Andrews and McIntyre (1976; 1978). The Eliassen-Palm (EP) flux is included in this formulation. Its magnitude represents the wave activity and its direction shows the direction of the group velocity of the wave (ex. Edmon *et al.*, 1980). In this paper wave-mean flow interactions around the SSB passages and their correlations to the reported responses are analyzed using both the transformed description by Andrews and McIntyre and the refractive index by Matsuno (1970).

2. Data and formulation

2.1 Data

The data used in this study are almost the same with those by both M81 and M83. Zonal Fourier components of geopotential height and temperature are obtained from the NMC gridpoint data at 10 pressure levels; 850, 700, 500, 300, 200, 150, 100, 50, 30 and 10 mb. They cover from 20°N to the North Pole at 5° latitude intervals for the period of 1964 to 1982. The data below the 100 mb level are at 12 hour intervals, but those above the 50 mb level are at 1200 GMT at 24 hour intervals.

The data are arranged to be data of 12 hour intervals by interpolation with the quad-

ratic function. Dates of the SSB passages are the same as M83. Those after 1979 are obtained by satellites observations. The method of superposing every SSB event on a key data is the same as M83.

2.2 Formulation

Following Palmer (1981, 1982) and Kanzawa (1982), the TEM formulation are expressed by,

$$\frac{\partial \bar{u}}{\partial t} - f \bar{V} = DF \quad (1)$$

$$\frac{\partial \bar{\theta}}{\partial t} = -\bar{\theta}_z \bar{W} + \bar{D} \quad (2)$$

$$-\frac{1}{\cos \phi} \frac{\partial}{\partial y} (\cos \phi \bar{V}) + \frac{1}{p} \frac{\partial}{\partial z} (p \bar{W}) = 0 \quad (3)$$

where

$$\bar{V} = \bar{v} - \frac{1}{p} \frac{\partial}{\partial z} \left(p \frac{\overline{v'\theta'}}{\bar{\theta}_z} \right) \quad (4)$$

$$\bar{W} = \bar{w} + \frac{1}{\cos \phi} \frac{\partial}{\partial y} \left(\cos \phi \frac{\overline{v'\theta'}}{\bar{\theta}_z} \right) \quad (5)$$

$$DF = \frac{1}{p \cdot a \cdot \cos \phi} \nabla \cdot \mathbf{F} = \frac{1}{p \cdot a \cdot \cos \phi} \cdot \left\{ \frac{1}{\cos \phi} \frac{\partial}{\partial y} (\cos \phi F_y) + \frac{\partial}{\partial z} (F_z) \right\} \quad (6)$$

$$\mathbf{F} = (F_y, F_z) = p \cdot a \cdot \cos \phi \left(-\overline{u'v'}, f \frac{\overline{v'\theta'}}{\bar{\theta}_z} \right) \quad (7)$$

(see Appendix A for undefined terms).

$\bar{V}$ and $\bar{W}$ are the residual mean meridional circulation and $\mathbf{F}$ is the Eliassen-Palm (EP) flux. This definition of EP flux is simple even though the dimension is not correctly written. If this EP flux divided by RT_s it is EP flux for unit volume (Kanzawa, 1982), and if by p_s it is EP flux for unit mass (Palmer, 1982).

The relation between DF and the residual mean meridional circulation is expressed by

$$\left(f^2 \frac{\partial^2}{\partial z^2} + N^2 \frac{\partial^2}{\partial y^2} \right) \varphi = f \frac{\partial DF}{\partial z} + \frac{\partial}{\partial y} \bar{D} \quad (8)$$

where φ is a stream function for $(\bar{V}, \bar{W})$. This equation indicates that a residual mean meridional circulation appears when $\partial DF / \partial z \neq 0$.

In this study, the basic field is defined as

a time average over the range of ± 15 days before and after each SSB passage. Hereafter, a basic field of a physical parameter A will be denoted by $[A]$ and a deviation from it by A° . In the first approximation, the following relations should be hold

$$-f[\bar{V}] = [DF] \quad (9)$$

$$-[\bar{\theta}_z \bar{W}] + [\bar{D}] = 0 \quad (10)$$

$$\frac{1}{\cos \phi} \frac{\partial}{\partial y} (\cos \phi [\bar{V}]) + \frac{1}{p} \frac{\partial}{\partial z} (p [\bar{W}]) = 0 \quad (11)$$

Hence eqs. (1), (2), (3) and (8) are rewritten as

$$\frac{\partial \bar{u}}{\partial t} - f \bar{V} = \frac{\partial \bar{u}^\circ}{\partial t} - f \bar{V}^\circ = DF^\circ \quad (12)$$

$$\frac{\partial \bar{\theta}}{\partial t} = \frac{\partial \bar{\theta}^\circ}{\partial t} = -(\bar{\theta}_z \bar{W})^\circ + \bar{D}^\circ \quad (13)$$

$$\frac{1}{\cos \phi} \frac{\partial}{\partial y} (\cos \phi \bar{V}^\circ) + \frac{1}{p} \frac{\partial}{\partial z} (p \bar{W}^\circ) = 0 \quad (14)$$

$$(f^2 \frac{\partial^2}{\partial z^2} + N^2 \frac{\partial^2}{\partial y^2}) \varphi^\circ = f \frac{\partial DF^\circ}{\partial z} + \frac{\partial}{\partial y} \bar{D}^\circ \quad (15)$$

These equations indicate that the time variation can be estimated by deviations from the basic field.

Unless otherwise stated, our results are based on 43 passages which occurred in 8 winters from 1964 to 1972. Basic mean fields of this study were obtained by superposition of these 43 cases, which are discussed in Appendix B.

Furthermore, to examine the property of wave propagation, we will use 'refractive index' by Matsuno (1970);

$$Q_k = \frac{\bar{q}_y}{\bar{u}} - \frac{k^2}{(a \cdot \cos \phi)^2} - \frac{f^2}{4N^2 H^2} \quad (16)$$

where

$$\begin{aligned} \bar{q}_y = & \frac{2\Omega \cos \phi}{a} - \frac{\partial}{\partial y} \left\{ \frac{1}{\cos \phi} \frac{\partial}{\partial y} (\bar{u} \cos \phi) \right\} \\ & + \frac{f}{p} \frac{\partial}{\partial z} \left(p \frac{\partial \bar{\theta}_y}{\partial z} \right) \end{aligned} \quad (17)$$

Following Palmer (1981), the relation between Q_k and F is,

$$\{f^2(F_y)^2 + N^2(F_z)^2\}^{1/2} = \frac{p_s}{2f} Z_k^2 \sqrt{Q_k} \quad (18)$$

where Z_k is the amplitude in zonal Fourier k -th component of the height field. F is guided along a ridge of Q_k/f^2 .

We also use the following relationship between DF and the eddy quasi-geostrophic potential vorticity (q') (Palmer, 1982)

$$DF = \bar{v}' q' \quad (19)$$

where

$$\begin{aligned} q' = & \frac{\partial v'}{\partial x} - \frac{f}{\cos \phi} \frac{\partial}{\partial y} \left(-\frac{\cos \phi}{f} u' \right) \\ & + \frac{f}{p} \frac{\partial}{\partial z} \left(p \frac{\theta'}{\theta_z} \right) \end{aligned} \quad (20)$$

as is given by Matsuno (1970). These equations are the TEM equations system.

3. Results

3.1 Zonal mean field

First the main results by M81 and M83 are summarized. Fig. 1 shows meridional cross sections of $\bar{u}^\circ$ for 12 days from Day -6 to Day 5 (where Day $n(-n)$ represents n th days after (before) the SSB passage). Characteristics of the time variations are listed below.

<U1>: In the vertical zone below the 30 mb level around 80°N , $\bar{u}^\circ$ is accelerated eastward in the period from Day -4 to Day -0.5 (not shown) but decelerated after the SSB passages. This phenomenon has two centers: One at the 300 mb level is named as <U1a> and the other at the 50 mb and 100 mb levels as <U1b>. As was already shown in Fig. 5 in M83, <U1a> holds the thermal wind balance to the temperature response which takes place in the polar troposphere. <U1b> is also a reflection of the tropospheric temperature response but is a little modulated by some temperature changes within the lower stratosphere.

<U2>: At the 10 mb level around 70°N , $\bar{u}^\circ$ is accelerated in the period from Day 1 to Day 3, and decelerated after that. This variation must be related with the temperature change reported as a SSB response by M81.

<U3>: At 40°N around the 300 mb level, the maximum of $\bar{u}^\circ$ occurs on Day -1.5 (not shown). This was pointed out in Fig. 4 in M83.

Fig. 1 Meridional cross sections of $\bar{u}^\circ$ for 12 days from Day -6 to 5. The interval of contours is 0.25 m/s. Eastward anomaly regions are shaded. The name of each region is written in the text (3.1).

3.2 Effects of waves on the zonal mean fields

Fig. 2 shows meridional cross sections of DF° for zonal wavenumbers (k) 1–3. DF° is connected to $\bar{u}^\circ$ through eq. (12). Waves shorter than those of $k=3$ are excluded because of their small contribution.

During 4 days from Day -5 to Day -2, DF° keeps positive in the northern part of 50°N in the 200–500 mb layer, this being named as <Div. 1>. The maximum of <Div. 1> occurs in the 300–500 mb layer between 65° and 70°N on Day -4.5 (see Fig. 4(a)). The maximum area moves upward from Day -2 to Day -1, this being named as <Div. 2>. Other regions where $DF^\circ > 0$ appear around the 10 mb level from Day -2 to Day 2 and in the 30–100 mb layer around Day -3. Those are named

<Div. 3> and <Div. 4>, respectively.

On Day 0, negative DF° appears at 80°N, this being named as <Con. 1> and one at 60°N in the 200–300 mb layer, as <Con. 2>. On Day 1.5 (not shown) at the polar region in the layer 20–300 mb is <Con. 3>. Though <Con. 2> disappears soon, both <Con. 1> and <Con. 3> remain until Day 3. In the following sub-sections, mutual relations of DF° , $\bar{u}^\circ$ and $\bar{T}^\circ$ will be discussed.

3.2.1 EP flux divergences and meridional circulation

<Div. 1> and <U1a> are grossly balanced with each other. However, the following two points seem to be unreasonable: One is that <U1a> is biased poleward from <Div. 1> and the other is that top of <U1a> spreads over

Fig. 2 Meridional cross sections of DF° for $k=1-3$ on the same days as Fig. 1. The interval of contours is 0.5 m/s/day and thick contour is used where deviation is exceed 2.0 m/s/day . The region where $DF^\circ > 0$ is shaded. The name of each region is the same as in the text (3.2).

that of $\langle \text{Div. 1} \rangle$. These can be explained as follows:

A schematic cross section of the residual mean circulation ($\bar{V}^\circ$, $\bar{W}^\circ$) expressing a wave mean flow interaction in the period of $\langle \text{Div. 1} \rangle$ is shown in Fig. 3. The A region is the center of $\langle \text{Div. 1} \rangle$. The eastward acceleration in the C region is greater than that in the B region, because $\bar{V}^\circ$ in the former is smaller than that in the latter. The westward acceleration due to the anticlockwise residual mean circulation is dominant in the D region. The temperature in the E region decreases because of the upward $\bar{W}^\circ$. The magnitude of $\bar{W}^\circ$, which is estimated from Fig. 4(a) by eq. (14) with assumption of $\bar{W}^\circ = 0$ at the tropopause, is almost the same as that of $\bar{W}^\circ$ estimated from temperature decrease in M83 by eq. (13) under assumption of $\bar{D}^\circ = 0$.

The equatorward $\bar{V}^\circ$ region in the 150–300 mb layer to the south of 40°N shown in

Fig. 3 A schematic meridional cross section of wave and flow activities in the period of $\langle \text{Div. 1} \rangle$. Thick line shows the residual mean circulation ($\bar{V}^\circ$, $\bar{W}^\circ$). The region where $DF^\circ > 0$ is hatched and enclosed by thin line and especially the region where $DF^\circ \gg 0$ is shaded. This region corresponds to $\langle \text{Div. 1} \rangle$. The $\langle \text{U1a} \rangle$ and $\langle \text{U3} \rangle$ region are also enclosed by thin line. The name of regions is seen in the text (3.2.1).

Fig. 4 Meridional cross sections of $f\bar{V}^\circ$ on Day -4.5, -3.5 and -2.5 (a) and on Day -0.5, 0.5, and 1.5 (b). This $\bar{V}^\circ$ is estimated as the residual term in eq. (12). The interval of contours is 0.25 m/s/day. The region where $\bar{V}^\circ > 0$ is shaded.

Fig. 4(a) is balanced with $\langle U3 \rangle$. The upward $\bar{W}^\circ$ in the F region connecting the southward $\bar{V}^\circ$ in high latitudes and that in low latitudes is supposed by the temperature descent in the 150–200 mb layer around 40°N (not shown). These show that all variations of $\bar{u}$ and $\bar{T}$ before the SSB passages is mutually related to $\langle \text{Div. 1} \rangle$. Moreover, the circulation pattern in Fig. 3 is just as the same as the reverse of the basic residual mean circulation shown in Appendix B. This implies that the basic residual mean circulation is weakened by $\langle \text{Div. 1} \rangle$.

$\langle \text{Div. 4} \rangle$ is expected to affect $\langle U1b \rangle$. However, any correlations of the position and the magnitude between $\langle U1b \rangle$ and $\langle \text{Div. 4} \rangle$ are not clear, merely because the center of $\langle \text{Div. 4} \rangle$ lies out of the frame (see 3.4.2). $\langle \text{Div. 2} \rangle$ spreads $\langle U1b \rangle$ to the south and $\langle \text{Div. 3} \rangle$ must create $\langle U2 \rangle$.

3.2.2 EP flux convergences

Fig. 4(b) shows the same cross sections as Fig. 4(a) except for those around Day 0.

Maxima of poleward $\bar{V}^\circ$ take place over those of $\langle \text{Con. 1} \rangle$ and $\langle \text{Con. 2} \rangle$. The circulation reverse to the circulation shown in Fig. 3 can be expected in polar and high latitudes. The downward part of this reverse circulation must be responsible for the increase of the temperature in the polar troposphere after Day 0 (M83).

$\langle U1a \rangle$ is cancelled out by $\langle \text{Con. 1} \rangle$, and $\langle U1b \rangle$ by $\langle \text{Con. 3} \rangle$. An interesting point is that the cancellation drifts $\langle U1a \rangle$ and $\langle U1b \rangle$ equatorward. In the case $\langle \text{Con. 1} \rangle$ and $\langle U1a \rangle$ are overlapped, the south part of $\langle U1a \rangle$ becomes a new $\langle U1a \rangle$ ridge, because the westward acceleration center due to $\langle \text{Con. 1} \rangle$ is biased poleward due to the meridional asymmetry of $\bar{V}^\circ$. In addition to that, the wide drift of $\langle U1a \rangle$ needs that of $\langle \text{Con. 1} \rangle$, which will be discussed in 3.4.3. As expected by eq. (15), $\langle \text{Con. 3} \rangle$ should result in upward $\bar{W}^\circ$ in the polar stratosphere, which is related to the temperature decrease there (M81). The phenomenon which counteracts $\langle U2 \rangle$ is not found in Fig. 2.

3.3 EP flux convergence and the VAI response

Both the position and time of <Con. 2> accord with those of descent of the VAI (W74). A relationship between DF° and the eddy geostrophic potential vorticity flux, as is indicated by eq. (19), is that the eddy potential vorticity is transferred to the south when $DF^\circ < 0$. The absolute vorticity used to define the VAI is closely related to the potential vorticity. Any modulation of the potential vorticity by DF° could be responsible for the VAI variation. Moreover, this modulation does not conflict with results from some researchers (Bhatnagar and Jakobsson, 1978; Williams, 1978; Williams and Gerety, 1980) using variations of the kinetic energy or the total squared vorticity.

3.4. Variations of EP flux

3.4.1 <Div. 1>

Fig. 5(a) shows meridional cross sections of F° and DF° for $k=1-3$, $k=1$, $k=2$, and $k=3$ on Day -4.5 when <Div. 1> is at the maximum. It is found that <Div. 1> consists of DF° for $k=2$ and $k=3$ and is independent of DF° for $k=1$. Fig. 5(a) shows that both the contributions are caused by decreases of F_z for $k=2$ and $k=3$ around Day -5 in the troposphere, which are shown as <P2> in Fig. 6(b) and <P3> in Fig. 6(c), respectively. <P2> of which the center is at 55°N in the 300-500 mb layer develops to Day -3.5, and <P3> of which the center is at 60°N below the 850 mb level develops to Day -4. Both phenomena create negative F_z° regions on the meridional plane, and this region moves up-

Fig. 5 Meridional cross sections of F° and DF° for $k=1-3$, 1, 2, and 3 on Day -4.5 (a), Day 0.5 (b) and Day 1 (c). The meridional (vertical) arrow at the middle upper-left represents a value of F_y (F_z) equal to $5.21 \cdot 10^{14}$ ($1.96 \cdot 10^{12}$) $\text{kg} \cdot \text{m} \cdot \text{s}^{-2}$.

Fig. 6 Meridional cross sections of normalized F_z for $k=1$ (a), $K=2$ (b), and $k=3$ (c). The normalized F_z is defined by dividing F_z by an average value of $[F_z]$ at each level. The interval of contours is 0.25, and positive regions are shaded. $\langle P1 \rangle$, $\langle P2 \rangle$ and $\langle P3 \rangle$ are discussed in the text.

ward into the stratosphere by wave propagations (shown in Fig. 6(b) and (c)). Therefore, $\langle \text{Div. 1} \rangle$ must be caused by the passage of this negative F_z° region through the upper troposphere, namely the transiency of the wave, because the value of $\partial F_z^\circ / \partial z$ at a confined level becomes positive in the time when the negative F_z° region approaches to there. This is expected from the fact that $\langle \text{Div. 1} \rangle$ disappears when the minimum of F_z caused by $\langle P2 \rangle$ and $\langle P3 \rangle$ reaches the tropopause (on Day -3 for $k=2$ and Day -2.5 for $k=3$).

3.4.2 Other divergent regions

$\langle \text{Div. 2} \rangle$ is caused by the passage of the negative F_z° region due to $\langle P2 \rangle$. In this case the contribution of $k=2$ is dominant, while F of $k=3$ tends to turn to the equatorward around the tropopause (see Appendix B).

$\langle \text{Div. 3} \rangle$ is produced by the passage of the negative F_z° region for $k=1$ in the first phase from Day -1.5 to 0. Fig. 6(a) shows that this structure is due to $\langle P1 \rangle$ that F_z for $k=1$ is reduced in the 100–150 mb layer around 70°N from Day -4 to Day -3 . After that

the negative F_z° for $k=2$ due to $\langle P2 \rangle$ reaches there and takes over the contributions of $k=1$ to $\langle \text{Div. 3} \rangle$ until Day 3.

$\langle \text{Div. 4} \rangle$ is probably due to $\langle P1 \rangle$, because waves in $\langle \text{Div. 4} \rangle$ comes through the $\langle P1 \rangle$ region (see Fig. 11(c) in Appendix B). However, this is not confident, because the meridional finite difference scheme used to calculate F_y° and $\partial F_y^\circ / \partial y$ have large errors in the polar region^{*1}.

3.4.3 $\langle \text{Con. 1} \rangle$

Fig. 5(c) shows that $\langle \text{Con. 1} \rangle$ is due to F for $k=1$ amplified and tilted the poleward in the troposphere. The increasing F for $k=1$ does not propagate into the polar stratosphere (not shown) and both maxima of $\langle \text{Con. 1} \rangle$ and that of F_z for $k=1$ occurs on the same data on Day 2. These facts suggest that $\langle \text{Con. 1} \rangle$ is mainly created by the dispersion of the wave.

This increase of F for $k=1$ can be diag-

^{*1} However, we can discuss about $\langle \text{Div. 1} \rangle$ and $\langle \text{Con. 1} \rangle$, because they are free from what meridional finite difference scheme is used.

Fig. 7 (a) Latitudinal distributions of Q_1 at the 300 mb level on Day -2, 0 and 2. (b) Same as (a), but for Q_2 at the 300 mb level. (c) Same as (a), but for Q_3 at the 500 mb level. The vertical scale is 10^{-12} m^{-2} . The number on the line shows day from the SSB passage. Thin line is the basic field.

nosed by Fig. 7(a) which shows the latitudinal structure of the refractive index for $k=1$ (Q_1) at the 300 mb level (About the calculation error of Q_1 in the polar region, see 3.4.5). Under normal conditions, Q_1 at the 300 mb level has the maximum at 70°N , where $\mathbf{F}$ for $k=1$ passes through from the troposphere to the stratosphere. Around Day 0, this maximum is moved poleward by the eastward wind ridge at 80°N corresponding to $\langle U1a \rangle$. Therefore, the direction of $\mathbf{F}$ for $k=1$ becomes poleward. Since Q_1 in the polar stratosphere is negative (not shown), the increasing $\mathbf{F}$ for $k=1$ in the polar troposphere is dispersed and creates $\langle \text{Con. 1} \rangle$. Together with the center of $\langle U1a \rangle$ drifting equatorward shown in 3.2.2, this maximum of Q_1 returns to 70°N and then the direction of $\mathbf{F}$ of $k=1$ turns to the south. This clearly suggests that $\langle \text{Con. 1} \rangle$ following $\langle U1a \rangle$.

3.4.4 $\langle \text{Con. 2} \rangle$

$\langle \text{Con. 2} \rangle$ is due to F_z for $k=2$ and $k=3$, as is seen in Fig. 5(b). The contribution of F_z for $k=2$ locates in the 200–300 mb layer and that for $k=3$ in the 300–500 mb layer. On Day -0.5, F_z for $k=2$ at the 300 mb level begins to increase, when that at the 200 mb level remains small (not shown). Its spatial variation gives rise to $\langle \text{Con. 2} \rangle$. Unlike $\langle \text{Con. 1} \rangle$, $\langle \text{Con. 2} \rangle$ disappears when F_z for $k=2$ reaches its maximum. As mentioned in 3.4.1,

this temporal relation suggests that $\langle \text{Con. 2} \rangle$ is generated by the transiency of the wave.

The variation of F_z for $k=2$ can be accounted for by Q_2 . Fig. 7(b) shows the latitudinal structure of Q_2 at the 300 mb level. Just before Day 0, the peak of Q_2 at 70°N disappears because of $\langle U3 \rangle$. After Day 0, this peak reappears and develops to Day 2 with some reduction of $\langle U3 \rangle$. The same change of Q_2 also occurs at the 500 mb level, but not at the 200 mb level (not shown). Therefore, F_z for $k=2$ increases below the 300 mb level and then $\partial F_z / \partial z$ become negative in the 200–300 mb layer. $\langle \text{Con. 2} \rangle$ is related only to the magnitude of Q_2 peak, though $\langle \text{Con. 1} \rangle$ is related to the position of Q_1 peak. Thus, the direction of $\mathbf{F}$ for $k=2$ is independent of $\langle \text{Con. 2} \rangle$: The dispersion of the wave is not great in $\langle \text{Con. 2} \rangle$ unlike $\langle \text{Con. 1} \rangle$, and the duration of $\langle \text{Con. 2} \rangle$ is shorter than that of $\langle \text{Con. 1} \rangle$ (see 3-2). The contribution of $\mathbf{F}$ for $k=3$ to $\langle \text{Con. 2} \rangle$ is the same as that of F_z for $k=2$. Q_3 at the 500 mb level behaves similarly as Q_2 (see Fig. 7(c)), but that at the 300 mb level does not (not shown), because the second term of Q_3 shown in eq. (16) is much larger than that of Q_2 in high latitudes. Therefore, in the case of $k=3$, the minimum of DF° occurs in the 300–500 mb layer.

3.4.5 $\langle \text{Con. 3} \rangle$

As already mentioned in 3.4.2, F_y in the

Fig. 8 Variations of t -value of F_z , of the amplitude of wave with respect to height (H) and temperature (T) for $k=1$ at the 150 mb level at 70°N (a), for $k=2$ at the 300 mb level at 60°N (b), and for $k=3$ at the 850 mb level at 70°N (c).

<Con. 3> region can not be estimated. Q_k in the <Con. 3> region also includes much larger errors generated by the meridional finite difference scheme than that in the <Con. 1> region, because the former is more biased to the pole than the latter and the meridional difference term of eq. (17), i.e., $\frac{\partial}{\partial y} \left\{ \frac{1}{\cos \phi} \frac{\partial}{\partial y} (\bar{u} \cos \phi) \right\}$, is dominant in the former, as is shown in Fig. 10 in Appendix B. Hence, it is meaningless to diagnose <Con. 3> by the distribution of Q_k .

4. Discussions

The dynamical processes after <P1>, <P2>, and <P3> have been discussed by use of the TEM equations system. We will discuss about <P1>, <P2>, and <P3> in the following. The magnitude of <P1>, <P2>, and <P3> in the 8 winters of 1964–1972 are extremely larger than other winters in the period of 1964–1982 (not shown). This indicates that the decrease of EP flux plays an essential role in the reported responses (W74, M81 and M83).

The refractive indices Q_2 and Q_3 in the troposphere around Day -5 (not shown) have not such variations as those around Day 0 shown in Fig. 7. This implies that <P2> and <P3> can not be diagnosed by the TEM equations system.

The statistical significance of these phenomena is analysed. Fig. 8 shows variations of Student's t value of F_z for $k=1$ at the 150 mb level at 70°N (a), for $k=2$ at the 300 mb

level at 60°N (b), and for $k=3$ at the 850 mb level at 70°N (c) around the SSB passages. These give an information about the statistical significance of <P1>, <P2>, and <P3>: <P1>, of which t value is -1.71, is statistically meaningless. <P3> of -4.24 is significant.

Fig. 9 Distributions of 3000 (DF° for $k=3$ and DF° for $k=2$) at 70°N in the 200–500 mb layer which constitute maxima of DF° for $k=2-3$ in the superposed result with ± 15 days interval. These superposes are performed by use of 43 key dates chosen randomly from the winters of 1964–1972 with more than 5 days intervals. 4 ellipses indicate 10%, 5%, 2.5%, and 1% level of χ^2 test respectively. The cross indicates the value of DF° for $k=2-3$ on Day -4.5. The EP flux used in this test is filtered by the same method as M83 to remove the seasonal trend.

$\langle P2 \rangle$ of -2.85 is hard to judge correctly from Fig. 8.

To study the significance level in more detail, we must use the so-called random sampling method (Haurwitz and Brier; 1981, and M83). Fig. 9 is a result for $\langle \text{Div. 1} \rangle$. It shows that $\langle \text{Div. 1} \rangle$ is beyond 99% confidence level, and both of the contribution of $k=2$ and 3 are effective for this confidence. Therefore, we should not exclude $\langle P2 \rangle$ which is the origin of the contribution of $k=2$ for $\langle \text{Div. 1} \rangle$.

Fig. 8 shows variations of t value of the amplitude of wave with respect to height (H) and temperature (T). These variations in $\langle P2 \rangle$ the $\langle P3 \rangle$ are much smaller than those of F_z . This fact is in accordance with the results by Williams and Gerety (1980) who found no significant variation in the kinetic and available potential energy of planetary waves. Therefore, $\langle P2 \rangle$ and $\langle P3 \rangle$ should be mainly responsible for the decrease of the westward inclination of wave phase for $k=2$ and 3 in the vertical. On the other hand, the variation in $\langle P1 \rangle$ accompanies a comparable variation of the kinetic energy.

5. Concluding remarks

In this study, the interactions between zonal mean fields and planetary waves around the SSB passages are described by the TEM equations system together with the refractive index. The conclusions of this study are:

1. The EP flux for $k=2$ and 3 in the troposphere had decreased during a few days from 5 or 6 days before the SSB passages.

2. This phenomenon coincides with the atmospheric responses reported so far (Wilcox *et al.*, 1974; Misumi, 1981; 1983).

3. However, this can not be diagnosed by the TEM equations system.

4. Not the change of the wave amplitude but the westward inclination of the wave phase to the vertical direction is responsible for the phenomenon.

5. There is a possibility that EP flux for $k=1$ in the stratosphere decreases before the SSB passages.

The two questions remain to be solved:

1. In the sense of the sun-weather rela-

tion, what can we consider from the fact that the response occurs on 5 days before the SSB passages?

2. Why does such a response occur only in the 8 winters limited from 1964 to 1972?

To approach these questions, Misumi (1986) has checked variations of some solar and interplanetary parameters around SSB passages, and found out that 10.7 cm solar radio flux intensity was smaller from 6 to 1 days before the SSB passages. This is a temporal variation limited in the same 8 winters as those used in this paper. This should be a key to the above questions. However, we do not easily understand a link between them. Any mechanism for that should be investigated in the future.

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Appendix. A

List of symbols.

(u, v, w), zonal, meridional, and vertical component of geostrophic wind velocity.

ϕ , latitude.

λ , longitude.

θ , potential temperature.

Ω , angular speed of rotation of the Earth.

H , constant scale height ($=7000$ m).

R , gas constant for dry air.

f , ($2\Omega \cdot \sin \phi$) Coriolis's parameter.

p , pressure in mbar.

p_s , surface pressure (1013.25 mbar).

a , radius of the Earth.

- D , diabatic heating.
 Φ , geopotential height.
 $\bar{X}$, zonal mean of physical parameter X .
 X' , deviation of physical parameter X from the zonal mean.
 N , $\left(= \left[\frac{g \bar{\theta}_z}{\bar{\theta}} \right]^{1/2} \right)$ Brunt-Väisälä frequency.

Appendix. B

The basic fields for the 43-case are discussed. These fields approximately correspond with the climatological mean field in the winters from 1964 to 1972, because they are the mean of 30 days around 43 key dates scattered all over 8 winters. Fig. 10 shows meridional cross sections of $[\bar{u}]$. It has good

Fig. 10 Meridional cross section of $[\bar{u}]$. The interval of contours is 2 m/s. The eastward wind is shaded.

accordance with other climatological means in winter. Fig. 11 shows $[\mathbf{F}]$ cross sections for $k=1-3$ (a), $k=1$ (b), $k=2$ (c) and $k=3$ (d). Some characters of these are listed as follows:

B-1. All waves are dispersed in the upper troposphere: The convergence center for $k=1$ is north of 80°N , that for $k=2$ at 75°N and that for $k=3$ at 60°N .

B-2. In the troposphere $[\mathbf{F}]$ for each of three wavenumbers is comparable, but $[\mathbf{F}]$ for $k=3$ is negligible small in the middle stratosphere, because most of it turns southward in the lower stratosphere.

B-3. The convergence or divergence of $[\mathbf{F}]$ must be equal to $-f[\bar{V}]$, see eq. (9). The pattern in Fig. 11(a) suggests the same residual mean circulation pattern as Fig. 6 of Edmon *et al.* (1980) and Fig. 3 of Karoly (1982), but the center of northward residual mean flow in the equatorial region seems to be biased to the north from that of their results. The maximum northward velocity in high latitudes is about 0.7 m/s.

It must be noted that after 1972, $[\mathbf{F}]$ for $k=1$ directs to the south in the upper-north part of Fig. 11(b), and so the divergent region locates there.

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Fig. 11 Meridional cross sections of $[\mathbf{F}]$ for $k=1-3$ (a), $k=1$ (b), $k=2$ (c), and $k=3$ (d). The meridional (vertical) arrow at middle upper-left represents a value of $F_y \cos \phi$ ($F_z \cos \phi$) equal to 2.09×10^{15} (7.84×10^{12}) $\text{kg} \cdot \text{m} \cdot \text{s}^{-2}$. The scale of the most thick arrow is 5 times greater than that of standard arrow, which is also 5 times greater than that of the thin arrow.

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変換したオイラー方程式系で見た太陽活動と気象の関係

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惑星間空間磁場境界通過に対する地球大気の応答として報告されている現象 (Wilcox *et al.*, 1974; Misumi, 1981, 1983) の原因と関連を調べた。惑星規模の力学過程に注目して屈折指数を含んだ変換したオイラー (TEM) 方程式を用いて解析した。以下 1964—1972 年間の北半球の冬における季節平均場からの偏差について議論している。

東西波数 2 と 3 成分の EP フラックスが通過 5 日前頃から高緯度対流圏において減少していた。この減少によって通過前 4.5~0.5 日間に高緯度の対流圏界面付近で E-P フラックスの発散が起り、次にこの発散は極域対流圏の温度を減少させ、又 Vorticity Area Index の減少を引き起こす。さらに波数 2 の EP フラックスの減少は、150 mb 面付近で通過 4 日前に始まる波数 1 成分の EP フラックスの減少と相まって成層圏極域における温度極少の原因になった。

以上の過程で EP フラックスの減少のみが TEM 方程式系で説明できない。特に対流圏内の EP フラックスの減少は統計的に有意でしかも他の応答と同じ期間にしか見出されない。

* 現在：仙台管区気象台