Observations of the Stratospheric Final Warmings in the Two Hemispheres

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Abstract

A comparison is made of the stratospheric final warmings which occurred in the two hemispheres. The dataset for this study consists of the NMC 1200 GMT analysis between 0.4 and 1000 mb during 1982. The transformed Eulerian mean diagnosis is used for examining the wave, mean-flow interaction.

The final warming occurred around March 31 in the Northern Hemisphere (NH) and around October 20 in the Southern Hemisphere (SH). Both warmings were associated with the enhanced planetary scale (wavenumber 1) wave activity. The final warming in the SH is more rapid and intense, which is consistent with the fact that the planetary scale wave activity in the SH stratosphere is more intense than that in the NH stratosphere during the spring season. In the SH the polar easterly did not descend below 10 mb after the final warming and circulation reversal below 10 mb occurred about one month later, while the polar easterly kept descending to 50 mb in the NH. The equatorial easterly in the lower stratosphere extended and connected to the polar easterly in the upper stratosphere of the SH, while the connection was not observed in the NH. A few days before the final warmings, the double jet structures were observed in the troposphere for both cases.

The feature of the final warming in the SH of 1982 is very similar to that of 1979 studied by Yamazaki and Mechoso (1985).

1. Introduction

The stratospheric sudden warming in the midwinter Northern Hemisphere (NH) has drawn great attentions since its discovery. Many studies have been performed for the midwinter NH warming (e.g., McIntyre, 1982 and references therein). As for the Southern Hemisphere (SH), Phillpot (1969) studied the stratospheric warming based on the conventional data. The analysis was restricted in the lower stratosphere and the major warming was not found in the SH. His analysis shows that the warming in the SH is more frequent in spring season than in midwinter. The early satellite measurements (Nimbus 3-5) provided the global information and our knowledge extended to the upper stratosphere (Fritz and Soules, 1970; Labitzke and Barnett, 1973; Barnett, 1975). It is found that major midwinter warmings, leading to a breakdown of the stratospheric circulation, take place only in the NH, although large sudden warmings take place even in the SH. In the SH, spring warmings seem to be more frequent and intense compared with midwinter warmings. A final warming (its definition in this paper will be given later) is important for the transport of the minor constituents such as ozone for both hemispheres, because an irreversible process is important for the net transport (Rood and Schoeberl, 1983). In this paper, the final warmings in the two hemispheres are studied.

Yamazaki and Mechoso (1985) investigated the Southern Hemispheric final warming which occurred in 1979. They showed that the sudden wind reversal from westerlies to easterlies occurred in the upper stratosphere in early October, 1979. However, the westeilies below 10 mb remained after the warming event. In this case, the easterlies in the low-latitude lower stratosphere seemed to play an important role. The purpose of this paper is to investigate whether these features observed in 1979 are the general ones in
the Southern Hemispheric final warming and to make a comparison between the two hemispheres.

Every year the stratospheric circulation changes from winter westerlies to summer easterlies. If the transition is gradual it is not called final warming in the present paper. To make a point clear, a tentative definition of the “final warming” is proposed. The transition must satisfy the following criterion for the minor warming at least. According to the WMO alert criterion, a stratospheric warming is called as “minor” if “a significant temperature increase is observed (i.e. at least 25 degrees in a period of a week or less) at any stratospheric level in any area of the wintertime hemisphere, measured by radiosonde or rocket-sonde data and/or indicated by satellite data.” Tentatively the threshold value for the temperature increase is set to be 20 degrees in the zonal mean, because the zonal mean is more convenient quantity for grid data. A stratospheric warming can be said to be “major” if “at 10 mb or below the latitudinal mean temperature increases poleward from 60 degrees latitude and an associated circulation reversal is observed” in addition to satisfying the criterion for the minor warming. The final warming is defined here as follows:

A significant temperature increase (i.e., at least 20 degrees in zonal mean in a period of a week or less) and circulation reversal to easterly are observed at any stratospheric level and this circulation remains after the event.

In late-February of 1979, an major wavenumber-2 sudden warming event occurred in the NH (Quiroz, 1979; Palmer, 1981) and several observational, theoretical and modeling studies have been made for this event. Although it occurred in late-winter, the circulation was reversed to westerlies after the event. Therefore, this intense warming does not satisfy the above criterion. However, the SH warming event in 1979 studied by Yamazaki and Mechoso (1985) satisfies the above criterion. In the SH, no final sudden warming was observed in 1980, 1981 and 1983. Circulation changes in these years are rather gradual and seen in November. In 1982, final warmings that satisfy the above criterion took place in both the NH and SH. As shown later, the final warming of 1982 in the SH is a minor warming and that in the NH is a major warming according to the WMO criterion. The 1982 final warmings are analyzed in this paper.

2. Data

The dataset for this study consists of the U.S. National Meteorological Center (NMC) 1200 GMT analysis during 1982. The meteorological variables available are temperature and geopotential height in the stratosphere. In addition, winds are used for tropospheric levels. Winds in the stratosphere are obtained from the geopotential fields by using the geostrophic relations. The 18 pressure levels from 0.4 through 1000 mb are used. Missing data are filled by interpolating linearly in time when necessary. Note that tropospheric data during the period from June 30 to July 17 are missing.

3. Evolution of the flow

3.1 Zonal mean fields

The time evolution of zonal mean zonal geostrophic wind at 2 mb for 1982 is shown in Fig. 1. Winds over 10°N–10°S were linearly interpolated from the northern and southern values. The figure has been plotted based on five day averages. Missing data are excluded from the average. All the figures that span one year are plotted in the same way. Numbers below the abscissa indicate the pentad numbers.

At mid-latitudes, both the winter westerlies and summer easterlies are stronger in the SH than those in the NH. In fall, the westerly jets start to form at high-latitudes and low-latitudes in both hemispheres. In early winter, the westerly jets are located at around 50°N/S in both hemispheres. In northern mid-winter, the jet shifts poleward followed by the sudden warming at the end of January and westerly is not strong during the rest of winter. The poleward shifting of the jet is also seen in the SH in mid-July, but the intensity of the jet remains strong even after the mid-July event. This different evolution of the stratospheric flow between hemispheres is shown by Shiotani and Hirota (1985). The shifting of the stratospheric westerly jet in the SH is also seen in other years (Harwood, 1975; Hartmann, 1976). Hartmann et al. (1984) showed the poleward shifting of the stratospheric westerly jet also oc-
Fig. 1. Latitude-time section of zonal mean zonal wind at 2 mb. Abscissa is a pentad number. Contour interval is 10 m/s. Negative values (easterlies) are shaded.

Fig. 2. Time evolution of zonal mean temperature at 10 mb in the Northern Hemisphere (top) and in the Southern Hemisphere (bottom). Thick solid lines denote values at 80°N and 80°S, dashed lines 40°N and 40°S, thin solid lines at the equator.
occurred in mid-July of 1979 in the SH. However, the shift is rather gradual and delayed in 1980 and 1981 (see Mechoso et al., 1985). It is intriguing that these years did not have significant final warmings in spring.

Rapid transitions from westerly to easterly are found at the end of March for the NH and in mid-October for the SH. After the transitions, easterlies weakened but continued throughout summer. Zonal mean temperature increases at 10 mb, 80°N/S exceed 20°K for both hemispheres (see Fig. 2). Thus, these events satisfy the criterion for the final warming mentioned in Section 1. Fig. 2 indicates that the temperature increase during the SH event is more rapid and intense than that during the NH event. The slight temperature decreases at 40°N/S are observed in both hemispheres. The temperatures at the equator show no significant changes during the final warmings. Exponential relaxations after the warmings are seen for both events. Vacillation of period of about 10 days is seen before the final warming in the SH.

3.2 Seasonal march of wave activity

Before analyzing the final warmings, let us briefly survey the seasonal evolution of wave activities and understand the final warmings in the context of seasonal change. To illustrate the seasonal evolution of waves, total amplitude of geopotential height waves for wavenumber 1 to 10 \( A_{1-10} \), is calculated as follows:

\[
A_{1-10} = \left( \sum_{i=1}^{10} A_i^2 \right)^{1/2}
\]

where \( A_i \) is the geopotential wave amplitude for wavenumber \( i \). In the stratosphere, wavenumbers 1 and 2 are dominant and smaller scale waves are negligible. Therefore, \( A_{1-10} \) is roughly equal to \( A_{1-2} \). Latitude-time plots of \( A_{1-10} \) at 10 mb for both hemispheres are shown in Fig. 3. In the NH, the maximum amplitude of about 1500 m is found in January. On the other hand, in the SH...
the maximum amplitude of about 1400 m is seen in October. Namely, the wave activity in the NH is most intense during mid-winter and that in the SH is most intense in late winter to early spring. The different behavior of the wave activity between hemispheres has been pointed out by Hirota et al. (1983) and Shiotani and Hirota (1985). The final warming in the SH took place during the most active season of the wave activity. On the other hand, the final warming in the NH took place during the decreasing phase of the wave activity. The maximum amplitude of about 1000 m during the final warming in the NH is less than that in the SH.

Fig. 4 shows the time-height sections of amplitudes of geopotential height wavenumber 1 at 65°N/S for both hemispheres. Shadings indicate the regions where the amplitudes are over 500 m.

In the NH, large amplitudes are seen in October in the upper stratosphere. The amplitude became strong and the center of wave activity shifted downward during early winter. The maximum amplitudes are found in the middle stratosphere (2–10 mb) during mid-winter to late winter. The amplitude associated with energy density, i.e., the one multiplied by square root of pressure \((A\cdot(P/P_0)^{1/2})\), has the maximum between 10–30 mb (figures are not shown). At the end of March, wave activity is enhanced and the maximum is located around 10 mb level. This enhanced wave activity corresponds to the final warming. After the final warming, wave amplitude are very small.

In the SH, large amplitude first appeared in July, became stronger and shifted downward toward October. Large amplitudes are seen in the upper stratosphere (above 2 mb) during winter. From late September to October, wave amplitude became very large and the maximum appeared at 5–10 mb level. The maximum of the normalized amplitude is seen at 30–50 mb level. The final warming in mid-October is related to this enhancement of wave activity. After the final warming, the wave amplitude is quickly de-

![Fig. 4. Height-time sections of geopotential amplitude of wavenumber 1 at 65°N (top) and at 65°S (bottom). Thick contour interval is 100 m. Thin contour denotes 50 m. Values greater than 500 m are shaded. Values less than 50 m are dotted.](image-url)
increased. Although wave structures during the final warmings of both hemispheres are similar, seasonal trends of wave activity are quite different each other. Similar figures for wavenumber 2 are made, but there is no amplification of the wave during the final warmings. The both final warmings are associated with wavenumber 1 waves.

3.3 Synoptic fields

In this subsection, synoptic maps are shown. Geopotential height fields at 2 mb and 10 mb for the NH are shown in Fig. 5. At 2 mb, the ridge was formed over the Aleutian region on March 28, 1982. The Aleutian high became strong during the following few days and occupied the polar region in April 3, 1982. The low belt at 50°–60°N surrounding the polar region is seen on April 3. At 10 mb, the Aleutian high strengthened and moved poleward during the period. This warming event clearly shows the wavenumber 1 wave activity.

The evolution of synoptic fields for the SH is shown in Fig. 6. Polar vortex has been strong and situated near the pole until October 18. On October 20, the high in the Australian sector suddenly strengthened and occupied polar region on October 22. A low belt surrounding the polar high is also seen on October 22. Wavenumber 1 pattern is dominant in the SH warming. Compared with

![Fig. 5. Geopotential height maps for the Northern Hemisphere on March 28, March 30, April 1 and April 3, 1982 (from left to right) at 2 mb (top) and 10 mb (bottom). Contour interval is 200 m.](image)

![Fig. 6. Geopotential height maps for the Southern Hemisphere on October 16, 18, 20, 22, 1982 (from left to right) at 2 mb (top) and 10 mb (bottom).](image)
the warming event in 1979 (see Fig. 4 of Yamazaki and Mechoso (1985)), the location of the high in this case is situated eastward of that in 1979. At 10 mb, the polar low is being weakened during this period and its location is displaced from the pole. The high, however, did not occupy the polar region. Thus, the zonal mean zonal wind at 10 mb in high latitude was still westerly even after the event, which is different from the NH.

After the warming events, the polar highs in both hemispheres shranked. However, the polar lows had never recovered and the synoptic fields gradually changed to the summer circulation pattern in both hemispheres. The time evolutions of both warming are similar except for the rapidity of the SH case at 2 mb.

3.4 Daily evolution of the zonal mean fields

Latitude-height sections of zonal mean temperatures and zonal winds are shown in Figs. 7 and 8 for the Northern and SH, respectively. Before the warming, polar jets exist in the middle stratosphere. In particular, the one in the SH is strong and its maximum exceeds 45 m/s. It is interesting that double jet structure is seen in the tropospheres of both hemispheres. In the NH, besides the subtropical jet, the maximum westerly is located at 47°N, 250 mb. In the SH, the maximum is located at 42°S, 250 mb. Their locations are quite similar.

Just after the warming events (middle panels of Figures 7 and 8), the easterly regions were formed above 10 mb and poleward of 50°N/S in both hemispheres. In the SH, the upper polar easterly is connected with the easterly in the equatorial lower stratosphere. This feature is similar to the 1979 warming event (Yamazaki and Mechoso, 1985). On the other hand, polar easterly in the NH is not connected to the equatorial easterly.

In the temperature fields, the warming in the polar middle stratosphere is the prominent feature in both hemispheres. In the SH upper stratosphere, the temperature decrease is noticeable. Note that vertical temperature gradient in high latitudes is much larger in the SH than that in the NH, i.e., the SH stratosphere is more stable than that in the NH.

In the troposphere, the double jet structures seen before the warmings are not found just after the events. Few days after the warming events (bottom panels of Figs. 7 and 8), strong westerlies appear at high latitudes (60–70°N/S) in the troposphere. In the stratosphere, the easterly region has progressed downward in the NH, while that retreated in the SH.

The evolutions of zonal wind and temperature are similar to the one predicted theoretically for sudden warmings (Matsuno, 1971; Holton, 1976). This implies that the final warming is physically the same phenomena as a typical sudden warming in mid-winter. The descent of the critical line after the warming event seen in the NH is similar to Matsuno’s (1971) results. On the other hand, the SH case is similar to Holton’s (1976) results with respect to the absence of the critical line descent.

4. Diagnostics of wave, mean-flow interactions

The transformed Eulerian mean equations (Andrews and McIntyre, 1976; Edmon et al., 1980; Palmer, 1981) are used to elucidate wave, mean-flow interactions. The Eliassen-Palm (E-P) flux vector defined by

\[ F = \rho_0 a \cos \phi \left( -\bar{u}\bar{v}' + f \bar{v}' \theta'/\bar{\theta} \right), \]  

represents the direction of wave energy propagation. In (4.1) \( u \) and \( v \) are the zonal and meridional components of the horizontal velocity; \( \theta \) is the potential temperature; \( z = -H \ln(p/p_s) \) where \( H \) is the scale height (7 km) and \( p_s \) is a reference pressure (1000 mb); \( \rho_0 (z) = \rho_s \exp(z/H) \) where \( \rho_s \) is a reference density; \( f \) is the Coriolis parameter; \( a \) is the radius of the earth; \( \phi \) is latitude. Overbars denote zonal averages while primes denote deviations from those averages. The divergence of the E-P flux vector, which is related to the northward potential vorticity flux, appears in the transformed Eulerian mean momentum equation

\[ \frac{\partial \bar{u}}{\partial t} - f \bar{v}^* - \bar{\tau} = \nabla \cdot F = (\rho_0 a \cos \phi) = DF, \]

where \( \bar{\tau} \) is the subgrid scale momentum source; \( DF \) is called wave forcing or wave driving. The meridional component of the residual mean circulation is given by
Fig. 7. Latitude-height sections of zonal mean temperature (left) and zonal mean zonal wind (right) on March 29, April 2, April 6, 1982 (from top to bottom) for the Northern Hemisphere. Contour interval is 5°K for temperature and 5 m/s for zonal wind. Negative values are shaded.
Fig. 8. Same as Fig. 7 except for October 17, 21, 25, 1982 and for the Southern Hemisphere.
Fig. 9. Eliassen-Palm cross-sections for the Northern Hemisphere. The periods are March 26–29 (top), March 30–April 2 (middle) and April 3–6 (bottom). In the left panels, contours denote wave forcing. Contour interval is 5 m/s/day. The length of E-P vectors in the left panels is multiplied by the factor exp (z/H). Right panels show only the direction of E-P vectors.
Fig. 10. Same as Fig. 9 except for the Southern Hemisphere. The periods are October 14–17 (top), October 18–21 (middle) and October 22–25 (bottom).
The transformed Eulerian mean thermodynamic equation is given by

$$\bar{\vartheta} = \vartheta - \frac{1}{\rho_o} \frac{\partial}{\partial z} (\rho_o \bar{v} \bar{\theta}'/\bar{\theta}_i)$$ \hspace{1cm} (4.3)

The vertical component of the residual mean circulation is given by

$$\bar{\omega} = \bar{\omega} + \frac{a \cos \phi}{\vartheta} \frac{\partial}{\partial \phi} (\cos \phi \cdot \bar{v} \bar{\theta}'/\bar{\theta}_i)$$ \hspace{1cm} (4.5)

where $\bar{Q}$ is the radiational heating rate.

A series of 4 day mean E-P cross sections during the warming period are shown in Fig. 9 for the NH and in Fig. 10 for the SH. In these plots, the length of E-P flux vectors has been multiplied by the factor $\exp(z/H)$, which is approximately equivalent to dividing by density (Mechoso et al., 1985). Without this factor, the E-P vectors in the stratosphere become too small to draw.

In the NH, the E-P vectors in the high-latitude stratosphere are directed upward during the period March 20 – April 6. Negative wave forcing, i.e., convergence of the E-P flux, has a large value at 0.4 mb during the first sub-period, though the accuracy of computation near the upper boundary is questionable. During the second sub-period, a large negative wave forcing as much as -20 m/s/day is found at 2 mb, 70°N. This value is roughly twice as large as the observed deceleration rate. During the third sub-period, the negative wave forcing is confined in the polar stratosphere below 10 mb. The pattern of the wave forcing during the second and third sub-period are similar to the observed deceleration in the stratosphere except for the upper most layer, though the wave forcing is larger than the observed one. This implies that the Coriolis term tends to compensate the wave forcing but is still smaller than the wave forcing term (see eq. (4.2)). Therefore the transformed Eulerian mean formulation is useful for the diagnosing the wave, mean-flow interactions. Since the second term on the left hand side of eq. (4.2) should be positive, $\bar{\sigma}^*$ is presumed to be poleward. Therefore the residual mean circulation in this case is presumably similar to that shown by Dunkerton et al. (1981). They used the transformed Eulerian mean diagnostics for the sudden warming simulated by the semi-spectral primitive equation model.

In the SH, the E-P vectors are nearly horizontal during the first sub-period in the middle and upper stratosphere. The wave forcing generally shows small positive value in high-latitudes, in accord with the observed slow acceleration during the sub-period. During the second sub-period, the E-P vectors become large and vertical in the stratosphere. Large negative wave forcings with the peak value of −40 m/s/day are computed about 2 mb, 70°S. Compared with the observed wind change, the wave forcing is again twice as large. Discussion on the residual mean circulation mentioned for the NH is also applicable to this case. During the third sub-period, the E-P vectors become quite small and wave forcing is positive in high-latitudes. In addition to the radiational effect, the wave, mean-flow interaction must be responsible for the observed acceleration of zonal flow after the event.

Looking at the troposphere, we find that the wave forcing is mostly negative. In particular, large negative values as much as −10 m/s/day are found in the upper troposphere (300–400 mb). Observed zonal wind changes are much less than the wave forcing. Compensation of the wave forcing by the Coriolis term and dissipation is more complete in the troposphere than that in the stratosphere. Nevertheless, Hartmann et al. (1984) showed that the correlation between wave forcing and observed zonal wind change is high in the upper troposphere. Therefore, the relative change of wave forcing will give us a useful information on the wave, mean-flow interactions. During the first sub-period, negative values of wave forcing over the region of 40°–60°N is small in the upper troposphere (200–400 mb). The wave forcing over this region has changed to be large negative in the next sub-period. This time change of wave forcing is consistent with the observed wind change. The change of wave forcing can be clearly seen in the E-P vector fields. The E-P vector in the equatorward of 40°N, which was large and directed equatorward during the first sub-period, has become weak during the second sub-period. The direction of E-P vectors in 40–50°N changed from equatorward to poleward. What is more, the upward flux in the lower troposphere in mid-latitudes become
strong. Thus the large negative forcing is formed in 50–60°N during the second sub-period. Ultra-
long waves (wavenumber 1–3) has little contribu-
tion to the wave forcing in the troposphere. Long waves are responsible for the wave forcing in the troposphere.

The E-P cross section for wavenumber one shows that the wave forcing due to wavenumber one is very small in the troposphere (figures are not shown). On the contrary, the wavenumber one contributes to most of the wave forcing in the stratosphere. In the NH, wavenumber one seems to have originated in the midlatitude tropopause region and have entered to the strato-
sphere. On the other hand, wavenumber one in the SH seems to have originated in the high-lat
itude troposphere and have propagated upward to the stratosphere.

5. Summary and discussion

A comparison of the stratospheric final warm-
ings occurred in the two hemispheres has been made. The results are summarized at Table 1. The date in Table 1 is defined as the first day when the zonal wind become easterly at 2 mb, 65°N/S. The timing of the final warming is late in the SH compared with that in the NH.

The warming rate in the SH is about 9°K/day at 80°S, 10 mb and about twice as large as that in the NH. Because the radiational heating rate is considered to be about 1°K/day or less, this warming is caused by descending motion of the residual mean circulation (see eq. (4.4)). Neg-
llecting the radiational term (Q), estimated down-
ward velocity at 10 mb, 80°N/S from the potential temperature cross sections is about 0.4 km/ day for the NH and 0.5 km/day for the SH. The estimated vertical velocity in the SH is not twice as large as that in the NH, because the stratifica-
tion in the SH is more stable than that in the NH.

It is an interesting question why the planetary wave activity in the SH stratosphere is maximum in spring. If we presume that the origin of plan-
tary wave activity in the stratosphere mainly lies in the troposphere, there are two possibilities. One is the planetary wave activity in the tropo-
sphere is intense in spring. The other is the zonal wind condition in spring is most favorable for the planetary waves to propagate from the tropo-
sphere to the stratosphere.

Let us discuss the first possibility. As seen in Fig. 4, the planetary wave activity in the SH tro-
oposphere is slightly larger in spring than that in midwinter. However, this enhancement of wave activity in spring is so small that this cannot fully explain the large seasonal change in wave activity.

<table>
<thead>
<tr>
<th>Date</th>
<th>NH</th>
<th>SH</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wind change</td>
<td>Maximum at 2 mb, 70°N</td>
<td>Maximum at 2 mb, 65°S</td>
</tr>
<tr>
<td></td>
<td>-8 m/s/day</td>
<td>-22 m/s/day</td>
</tr>
<tr>
<td>Temperature change</td>
<td>Maximum 10 mb, 80°N</td>
<td>Maximum 10 mb, 80°S</td>
</tr>
<tr>
<td></td>
<td>4°/day</td>
<td>9°/day</td>
</tr>
<tr>
<td>Easterly region</td>
<td>Descended to 50 mb.</td>
<td>Not descended below 10 mb.</td>
</tr>
<tr>
<td></td>
<td>Not connected to the equatorial easterly</td>
<td>Connected to the equatorial easterly</td>
</tr>
<tr>
<td>Wave forcing</td>
<td>Maximum at 2 mb, 70°N</td>
<td>Maximum at 2 mb, 70°S</td>
</tr>
<tr>
<td></td>
<td>-20 m/s/day</td>
<td>-40 m/s/day</td>
</tr>
<tr>
<td>Troposphere</td>
<td>Double jet before the warming.</td>
<td>Double jet before the warming.</td>
</tr>
<tr>
<td></td>
<td>Wavenumber one originated from middle-latitudes, propagated poleward.</td>
<td>Wavenumber one originated from high-latitudes, propagated upward.</td>
</tr>
</tbody>
</table>
shown in Fig. 3. E-P vectors of planetary waves are mostly horizontal and equatorward in middle-latitude upper troposphere during winter. Whereas, upward component of the E-P vectors becomes large in spring. This suggests that the second possibility is more plausible. Because zonal wind speed is very large (maximum speed exceeds 100 m/s) during winter, even the wave-number one cannot propagate into upper and middle stratosphere. Probably the moderate westerly in spring provides a favorable condition for planetary waves to propagate into the stratosphere. Other possibilities such as an in situ instability exist. Further study is needed to clarify this problem.

A final warming is also important for tracer gases, such as ozone. On October 28, 1982, a sudden increase of total ozone was observed at Syowa station (69°00′S, 39°35′E) on the Antarctica (Chubachi, 1984). About one month later, it was observed at Amundsen-Scott station (the South Pole). This phenomenon seems to closely related to the final warming studied in this paper. About one week delay of the ozone increase at Syowa station from the final warming is described synoptically as follows. As seen in Fig. 6, the shrunk polar low displaced from the pole toward the Indian Ocean Sector, where Syowa station is located, on October 22. On October 28, the polar low quickly reoccupied the South Pole. However, the polar vortex was much smaller and weaker than that before the final warming. Moreover, a high pressure is formed around 45°E, 45°S of which area extended to near Syowa station. Thus the sudden increase of ozone occurred on October 28 at Syowa station. However, the zonal mean field does not change significantly on October 28. Therefore the sudden increase of ozone on October 28 is considered to be a local phenomenon, but related to the final warmings.

As mentioned in this paper, the polar low still existed below 10 mb after the final warming in the SH. Complete destruction of the polar low in the lower stratosphere occurred one month later (late November). This is consistent with the sudden increase of ozone at Amundsen-Scott on November.

In the lower stratosphere, where the maximum density of ozone appears, the planetary scale eddy is responsible for vertical and horizontal transport of the ozone. Different behavior of the seasonal march in total ozone between two polar regions is consistent with the different behavior of the wave activity in the stratosphere.

The feature of the final warming in the SH during 1982 is very similar to that during 1979 studied by Yamazaki and Mechoso (1986). The poleward and downward shifting of the stratospheric westerly jet took place during July in 1979 and 1982. On the other hand, it took place during September in 1980 and during August in 1980 and 1983. In 1979 and 1982, the wave activity is more intense compared with that in 1980, 1981 and 1983 (see Mechoso et al., 1985 for 1979–82). There seems to be two kinds of the circulation evolutions during winter and spring in the SH. In active years (1979 and 1982), the shifting of the stratospheric westerly jet takes place in midwinter and the winter-to-summer transition takes place as a intense final warming in the upper stratosphere. In quiet years (1980, 1981 and 1983), the shifting of the stratospheric westerly jet takes place in late winter and the winter-to-summer transition is rather gradual. Shiotani and Hirota (1985) found the relation between the timing of the shifting and the phase of the equatorial quasi-biennial oscillation (QBO). The present analysis also suggested the importance of the equatorial critical line in the final warming. In active years, the phase of the QBO is usually easterly. The idea of Holton and Tan (1980) that the QBO modulates the extratropical circulation of the middle stratosphere can be applicable to the SH.

As for the NH, there is a tendency for a weak stratospheric westerly jet and enhanced development of height-wave I in early winter, which leads often to the development of a major midwinter warming, during the easterly phase of the QBO (Labitzke, 1982; Holton and Tan, 1982). However, no clear relation can be found for the timing of the final warming (Labitzke, 1982).
Note that her definition of the final warming is different from that in this paper. Further studies on the interannual variability of the stratospheric circulation for both hemispheres are needed.

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References
成層圏最終昇温の南北両半球の比較

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（気象研究所）

南北両半球で起こった成層圏最終昇温の比較を行なった。この研究で用いたデータはNMCの1982年の1日1回（12Z）の解析値で、層は0.4 mbから1000 mbまでの18層である。波と平均場の相互作用を調べるために変形オイラー平均方程式系を用いた。

1982年の最終昇温は北半球では3月31日頃、南半球では10月20日頃起こった。北半球と比べて南半球の最終昇温の方が急速で激しく、このことはプラネタリウ波の活動が春には南半球でより活発であることと符合している。南半球では極の東風域は最終昇温直後も10 mb以下には下がらず、下部成層圏まで東風となるのは11月下旬である。一方、北半球では東風域は最終昇温直後、50 mbまで下降する。また南半球では低緯度下部成層圏の東風がのびて極域上部成層圏の東風域とつながる。一方、このようなことは北半球では見られない。両半球とも最終昇温の数日前に対流圏で二重ジェット構造が見られる。

両半球の最終昇温とも波数1のプラネタリウ波の増大に伴なって起きている。北半球では波数1の波は極地域対流圏で生成され極方向に伝播して高緯度対流圏界面に達し成層圏に入ったらと見られる。他方、南半球では波数1の波は高緯度対流圏で生成され上方へ伝播し成層圏へ入ったように見える。

1982年の南半球成層圏最終昇温は、Yamazaki and Mechoso (1985)によって調べられた1979年の最終昇温と非常によく似ており、南半球の波の活動が活発な年の特徴的な振舞いを示していると思われる。