Poleward Jet-Shift in the Southern Hemisphere Winter Simulated with a General Circulation Model

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Abstract

An investigation was conducted of the poleward jet shift appearing in the Southern Hemisphere winter, by means of data simulated by a general circulation model. The poleward jet-shift in the model is found to be well simulated, although the jet in mid-winter exists fairly southward compared with observation. It is conjectured that the poleward jet-shift in the model was preconditioned by the wave breaking of a forerunning Rossby wave with zonal-wavenumber 1, and then it occurred by the breaking of another strong Rossby wave packet with zonal-wavenumber 1 at the jet axis in the upper stratosphere.

1. Introduction

The poleward shift of the stratospheric polar night jet is an event that is known to occur in late winter of the Southern Hemisphere. In mid winter, the polar night jet core exists at mid latitudes of the upper stratosphere. As a seasonal march, however, the jet quickly shifts southward and downward (with a timescale of 10 days), accompanied by warming of the upper stratosphere.

No major warming occurs in the Southern Hemisphere winter, because the amplitude of the

planetary wave in the troposphere is smaller and the wind in the stratosphere is stronger than those of the Northern Hemisphere winter. A weak planetary wave will have a large enough amplitude at the upper-stratosphere, however, and then it takes mean flow interaction there. This is the jet-shift event in the upper-stratosphere of the Southern Hemisphere winter (Hartmann *et al.*, 1984; Mechoso *et al.*, 1985). The value of the zonal-wind change is comparable to that of the sudden warming in the Northern Hemisphere winter. After the shift, the activity of eastward-propagating Rossby waves with wavenumber 2 begins to be enhanced, and a variation of the jet strength occurs, and the jet finally disappears in mid to late spring in final warming. The jet-shift event is interesting because it is almost always a one-way event and it occurs at different times in different years. It is also known

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Fig. 1 Time-latitude cross sections of the (a) observed zonal-mean zonal wind and (b) the zonal-mean temperature at 1 hPa in the Southern Hemisphere winter from July to August of 1988. The contour intervals are 10 m/sec for (a) and 2 K for (b).

that final warming occurs early and the ozone hole becomes weak when the jet-shift occurs early. In contrast, if the jet-shift occurs late, the final warming becomes late and the ozone hole event starts dragging. In this sense, it is important to clarify the mechanism of the jet-shift event.

Few works concern the jet-shift event in the Southern Hemisphere winter, in contrast to sudden warmings of the Northern Hemisphere winter. Among them are Hartmann *et al.* (1984), Mechoso *et al.* (1985), and Shiotani *et al.* (1993) for observational studies. Hartmann *et al.* (1984) showed that the jet-shift in July 1979 was caused by planetary waves with zonal wavenumber 1 as well as the eastward traveling wave with wavenumber 2. Mechoso *et al.* (1985) showed that the jet-shift occurs in association with a burst of upwardly propagating planetary waves and the seasonal variation in thermal forcing. Based on 10-year statistics from 1980 to 1989, Shiotani *et al.* (1993) pointed out that the polar-night jet in the Southern Hemisphere winter is categorized into low-latitude jet and high-latitude jet years and they are intimately correlated to the wave activities for each year. These works show some typical characteristics about the jet-shift, but are not sufficient to understand the mechanism.

Here, we shall show typical example of the jet-shift event in the real atmosphere. Figure 1 shows (a) the observed zonal-mean zonal wind and (b) temperature, at 1 hPa from July to August of 1988, which have been analyzed from NMC stratospheric data. The zonal wind is calculated from the non-linear balance wind relation as described by Randel (1987). The polar-night jet existed about 40° S until early-July and it gradually shifted to 60° S in late-July. We found that the jet-shift accompanied a

warming of about 14 K 20 July. The warming had a peak value of about 16 K around 2 hPa.

In this paper, we present an example of the poleward jet-shift appearing in a 10-year simulation of the Meteorological Research Institute general circulation model (MRI-GSPM). We have also examined a possible mechanism for the poleward jet-shift in the Southern Hemisphere winter through the analysis of simulated data.

In the next section, we will discuss the features of the poleward jet-shift simulated in the model. Section 3 describes the possible mechanism for the poleward jet-shift in the model by means of the Eliassen-Palm (EP) flux analysis. Section 4 centers on conclusions and remarks.

2. Poleward jet shift simulated with the general circulation model.

2.1 Outline of the GCM and the data

The GCM used in this study is the R24L23 version of the MRI-GSPM. The model has 23 vertical layers, extending from the surface to 0.05 hPa and, horizontally, uses spectral transformation at a rhomboidal truncation of 24. Standard physical processes are included. The effect of gravity-wave drag is simply expressed by Rayleigh friction. The diurnal cycle of solar radiation is explicitly calculated, as is the seasonal cycle. The details and performance of the R13L23 version can be found in Shibata and Chiba (1990). The analysis of the tide and the normal mode Rossby waves in the R24L23 version of the model can be seen in Chiba and Shibata (1992), Kuroda *et al.* (1994), and Kuroda and Chiba (1995).

A 10-year simulation was made using the R24L23 version of the model (Chiba *et al.* 1996). Use

Fig. 2 Time-latitude cross section of (a) the zonal-mean zonal wind and (b) the zonal-mean temperature at 1 hPa in the Southern Hemisphere winter from July to August of GCM simulated data. Contour intervals are 10 m/sec for (a) and 2 K for (b). Negative values are shaded.

Fig. 3 Latitude-height cross section of the zonal-mean zonal wind velocity (a) before the poleward jet-shift (16 July) and (b) after the jet-shift (26 July). The contour intervals are 10 m/sec for both figures. Negative values are shaded.

is made of data simulated from the 6th model year when the poleward jet-shift clearly occurred in the Southern Hemisphere in late July.

2.2 Brief description of the poleward jet-shift

Figure 2(a) is the time-latitude cross section of the zonal-mean zonal velocity in the Southern Hemisphere at 1 hPa from July to August. The axis of the polar night jet exists at 60°S until mid-July. However, from the beginning of July, the northern envelope of the jet gradually reduces its velocity, and the axis shifts to 70°S in late-July. Figure 2(b) is the same as Fig. 2(a) except for temperature. Warming of about 15 K occurs in late July. This value is comparable to the observed values in the real atmosphere (Hartmann *et al.* 1984 and Fig. 1(b)).

Figures 3(a) and (b) show the latitude-height

cross section of the zonal-mean zonal velocity on 16 July, and on 26 July. The axis of the jet shifted poleward and downward during this period. Figures 4(a) and (b) show the horizontal wind map on 16 July, and on 26 July at 1 hPa. The jet-shift occurs mainly due to the poleward shift of the strong wind region which existed south of Australia. This situation is also seen in observation.

This jet-shift in the GCM simulation is approximately similar to that occurs in the real atmosphere. In the real atmosphere, however, the polar night jet exists at latitudes from 40 to 50 degrees in mid-winter (Fig. 1(a)). In this simulation, we cannot simulate well the location of the polar-night jet in mid winter. Cold bias of the polar upper-stratosphere is also found in the model (compare Figs. 1(b) with 2(b)), and these are closely related by geostrophic wind relations. Systematic errors of these

Fig. 4 Horizontal distribution of the wind (a) before the poleward jet-shift (16 July) and (b) after the jet-shift (26 July) at 1 hPa. Contours indicate wind speed. Contour intervals are 10 m/sec for both figures.

types exist in many GCMs (e.g. Boville 1995, Hamilton 1995 and Pawson *et al.* 1995). In contrast, after the jet-shift, the location of the jet, and the vacillation of the wind are well simulated.

3. Analysis of the poleward jet-shift in the model

The jet-shift event is thought to be caused mainly by the activities from below of forced Rossby waves. In fact, we found that the period of the jet-shift event corresponds well to that of the amplification of quasi-stationary forced Rossby waves, as the observation by Hartmann *et al.* (1984) shows. Wave activity during this episode will be analyzed in terms of the Eliassen-Palm flux (EP flux) and the quasi-stationary refractive index to diagnose the ability for the propagation of the quasi-stationary wave in the mean field. The possible mechanism for the jet-shift summarized in Fig. 5 will be discussed in the following.

Figure 6 shows the vertical propagation of waves with wavenumber 1 and their effects on the zonal-mean zonal wind. Thin solid lines show the vertical component of the EP flux in an arbitrary

scale and thick solid lines show the zonal-mean zonal velocity. These are averaged for the latitude from 40°S to 60°S. On 3 July, the vertical component of the EP flux reached its maximum in the troposphere, but little affected the zonal wind in the stratosphere. On 14 and 16 July, twin peaks of the EP flux appeared at 850 hPa, and could be traced to the stratosphere. Here, the dash-dotted line shows the 1st packet, and the dashed line shows the 2nd packet in Fig. 5. It can be seen that the 1st packet decayed at about 10 hPa before the jet-shift event. Consecutively the 2nd packet propagated upward, causing the jet-shift.

The radiation of forced Rossby waves was mainly due to the formation of blocking around the Antarctic Continent. Figure 7(a) shows the geopotential height at 500 hPa when the vertical component of the EP flux has the large value (15 July). There appear strong ridges at the northward of Antarctic peninsula and the south Pacific ocean.

Possible Jet-Shift Mechanism

Fig. 5 Flowchart for the possible mechanism for the poleward jet-shift event simulated in the model.

Fig. 6 Temporal variation of the EP flux vertical component of the stationary wave with wavenumber 1 (thin line) and zonal-mean zonal velocity (thick line) at various levels from 850 hPa to 1 hPa. They are averaged from 40°S to 60°S. The dash-dotted line (dashed line) indicates the 1st packet (2nd packet) in the flowchart in Fig. 5.

Fig. 7 Horizontal distribution of (a) the geopotential height and (b) the vertical component of the 3D EP-flux on 15 July when the large EP flux radiated upward at 500 hPa. Contour intervals are 100m for (a) and $2 \times 10^5 \text{ kg/sec}^2$ for (b). Negative values are shaded for (b).

We shall call this pattern double blocking. It can be seen that heat flux formed by blocking makes the large vertical EP flux component. In fact, the 3-dimensional extension of the EP flux (Plumb 1985) (Fig. 7(b)) shows that the vertical component of the EP flux with wavenumber 1 is large around 55°S in this pattern. We found that this double blocking pattern occasionally appears around the Antarctic Continent. The flow pattern tends to fluctuate around this double blocking pattern, which has a large vertical EP flux component. The fluctuation of this flow pattern will be due to the instability of the flow itself. This double blocking pattern seems to be very intimately related to the shape of the Antarctic Continent itself.

The Rossby wave, in general, cannot propagate efficiently to the stratosphere unless a jet forms near the tropopause at around 70°S in addition to the subtropical jet. We call this pattern a double jet. The reason why Rossby waves can propagate efficiently in the double jet situation is that there forms a high refractive index region northward of this jet. This jet attracts Rossby waves to the stratosphere. If this mechanism does not operate, the tropopause becomes a barrier (Chen and Robinson 1992).

In mid winter, Rossby waves tend to propagate equatorward because of the high refractive index region at the equatorward at this time. The Rossby wave breaks in the low latitude region and the EP flux converges there. This is the reason why the equatorward region wind speed of 30°S gradually reduced until mid-July (Fig. 2(a)).

When the 1st packet propagates to the middle

stratosphere, it also breaks at a low latitude. Figure 8(a) shows the refractive index squared when the 1st packet propagates to the stratosphere and breaks at around 10 hPa. In this figure, we have shaded high-refractive-index regions whose square exceeds 20. Figure 9(a) shows the EP flux and wave driving (the acceleration of the zonal-mean zonal wind due to the divergence of the EP flux) at that time. The amplitude of the 1st packet is fairly large, largely decelerates the zonal wind at a low latitude, and accelerates in the high latitude around 5 hPa. According to the zonal wind change, a part of the large refractive index region shifts poleward. This leads the Rossby wave to propagate into the high-latitude region.

The 2nd packet propagates upward to the high-latitude region through the high refractive index region. Figure 8(b) shows the refractive index squared when the 2nd packet propagates upward. The 2nd packet propagates into the jet-axis region, finally saturating and breaking down there and making a large EP flux convergence. Figure 9(b) shows the EP flux and wave driving at that time. This EP flux destroys the jet core and, at the same time, drives residual circulation and causes a warming there. Figure 10 shows the residual circulation and the variation of the zonal-mean temperature 2 days after the date of Figure 9(b) when the largest warming occurs. Poleward and downward residual winds occur in the jet-axis region. The downward residual wind causes warming at around 70°S and 1 hPa. At the same time, a poleward residual wind provides the Coriolis acceleration of the zonal wind

Fig. 8 Latitude-height cross section of the refractive index square of the quasi-stationary wave with zonal-wavenumber 1 (a) when the 1st packet in Fig. 5 propagates into the low-latitude region (18 July) and (b) when the 2nd packet in Fig. 5 propagates into the jet axis region (24 July). The shaded region denotes a region whose refractive index square value exceeds 20.

Fig. 9 Latitude-height cross section of the EP flux vector (arrow) and the wave driving due to the divergence of the EP flux (contour) at the same time as in Fig. 8. The EP flux vector is magnified by the inverse root of the pressure. The contour intervals are $5 \times 10^{-5} \text{ m/sec}^2$, and negative values are shaded.

at the high-latitude region which is larger than deceleration due to EP-flux convergence. This causes the poleward jet-shift event.

When the wave with zonal wavenumber 1 changes the zonal-mean zonal wind, the waves with zonal wavenumber 2 appear. Figure 11 shows a Hovmoeller diagram of the geopotential height of the eddy, averaged from 60° S to 70° S at 1 hPa. Waves with zonal-wavenumber 2 are amplified during the jet-shift period. (Note that the ridge and trough lines approach at the jet-shift period.) This wave is created by the instability of the mean field, as was pointed out by Hartmann *et al.* (1984). This wave also contributes to the total change of the zonal-mean zonal wind. Figures 12(a) and (b) show wave driving with zonal-wavenumber 1 and from 1 to 5. We found that the wave driving due to the wave with zonal-wavenumber 2 shows a dipole pattern, which accelerate around 50° S and decelerate around 70° S in late July. This contribution enlarges deceleration due to that of zonal-wavenumber 1 at higher latitudes and reduces it at lower latitude (Fig. 12(b)). This negative wave driving and the residual velocity it drives is thought to cause the jet-shift.

4. Conclusion and Remarks

We succeeded in simulating the poleward jet-

shift event that occurs in the Southern Hemisphere winter with a general circulation model. Although we could not well simulate the location of the jet axis in mid-winter by the model, the jet in the model shifts from 60° S to 70° S in late winter.

We have proposed a possible mechanism for the jet-shift, being caused mainly by the activities of continually propagating forced Rossby waves from below (Fig. 5). In winter in the Southern Hemisphere, the main source of forced Rossby waves near the troposphere is the formation of the double blocking flow around the Antarctic Continent. When the zonal-mean zonal wind forms the double jet, Rossby waves propagate upward to the stratosphere. At first, the high refractive index exists at a low-latitude region, and waves propagate equatorward. When the packet of the large amplitude wave propagates upward, the packet collapses, causing the wind profile, then changes the high refractive index region to a high latitude. This event is thought to be preconditioning for the jet-shift. When large-amplitude Rossby waves continually propagate upward, they enter the jet-axis region directly. The breakdown of this Rossby wave is thought to cause the jet-shift event.

Hartmann *et al.* (1984) and Mechoso *et al.* (1985) have explained the direct origin of the shift

Fig. 10 Latitude-height cross section of the residual velocity (arrow) and the variation of the zonal-mean temperature (contour). The contour intervals are 1K/day, and negative values are shaded.

Fig. 11 Hovmoeller diagram of the geopotential height of the eddy, which is averaged from 60° S to 70° S at 1 hPa from July to August. Contour intervals are $2 \times 10^2 m$, and negative values are shaded.

Fig. 12 Time-latitude cross section of wave driving due to the divergence of the EP flux by (a) the wave with zonal-wavenumber 1, and (b) that with 1 to 5 in the Southern Hemisphere winter from July to August at 1 hPa. The contour intervals are $1 \times 10^{-4} \text{ m/sec}^2$.

by the formation of the dipole pattern of wave driving, which accelerates zonal wind in high latitude and decelerates in low latitude appearing in observed data. The dipole pattern of wave driving is not so clearly seen in Figures 12(a) and (b). We have considered the direct origin of the jet-shift as the residual wind of wave driving by this reason. The direct mechanism for the jet-shift may differ in nature and our simulation at this point.

The scenario presented here was made possible through simulation. It is not, however, verified by observation, so future works are needed that include comparison with observation.

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大循環モデルでシミュレートされた冬季南半球のジェットシフトについて

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冬季に南半球に現れるジェットの極方向へのシフトを大循環モデルのシミュレーション結果を用いて調べた。その結果、モデル中での真冬でのジェットの位置は観測に比べて南よりではあったが、ジェットのシフトがうまく表されている事例が見いだされた。

我々はシミュレートされたジェットのシフトの解析についても行ってみた。その結果、モデルでのジェットのシフトは、先行する波数1のロスビー波の碎波によってプレコンディショニングされた後、波数1のパルス波が上部成層圏のジェット軸で碎波する事によって起こることが分かった。

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