Seasonal Cycle of the Climatological Stationary Waves in the NCEP–NCAR Reanalysis

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ABSTRACT

The maintenance mechanisms of the climatological stationary waves and their seasonal cycle are investigated with a linear stationary wave model and the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis data from 1985 to 1993. The stationary wave model is linearized about the zonal-mean flow and subjected to the zonally asymmetric stationary wave forcings. It has rhomboidal wavenumber 30 truncation and 14 vertical sigma levels. The forcings for the linear model include diabatic heating, orography, stationary nonlinearity, and transient vorticity and heat flux convergences. The NCEP–NCAR reanalysis provides a high quality global dataset for this study.

When the linear model is subjected to all forcings, it reproduces reasonably well the climatological stationary wave seasonal cycle. The linear stationary wave theory is quantitatively valid at the upper-tropospheric levels for all months and the lower-tropospheric levels for the northern summer months (with pattern correlation greater than 0.8). At the middle- and lower-tropospheric levels for most of the months, the stationary wave theory is qualitatively valid (with pattern correlation greater than 0.5). The effect and relative importance of each individual forcing mechanism in maintaining the stationary waves and their seasonal cycle are determined by the linear model. Within the linear model framework, the global diabatic heating is found to be the most dominant forcing mechanism for the climatological stationary waves throughout the seasonal cycle. Subsequently, the seasonal cycle of the stationary waves is largely caused by the seasonal fluctuations of the atmospheric heating field. By comparison, the linear effect of orography is of less importance in both the Tropics and the extratropics. The effect of stationary nonlinearity is to modify the spatial structure of the stationary waves, particularly over extratropical North America. Comparatively, transient forcing has little contribution. By separating the tropical and the extratropical heatings in the linear model, it is found that the local thermal forcing has the dominant contribution to the local stationary wave seasonal cycle.

The relative contribution of the seasonally varying zonal-mean basic state and the seasonally varying forcing fields is also examined using the linear model. The seasonally varying zonal-mean basic state can account for the zonal-mean amplitude fluctuation of the stationary waves in the Tropics, as well as the seasonal change of the stationary wave spatial structure from September to May. It fails to capture the amplitude fluctuation of the Northern Hemisphere extratropical stationary waves and the northern summer stationary wave spatial structure. On the other hand, the effect of the seasonally varying forcing accounts largely for the zonal-mean amplitude fluctuation of the stationary waves in the Northern Hemisphere extratropics, as well as the transition to the northern summer stationary wave regime.

1. Introduction

The climatological stationary waves, or the long-term mean of zonally asymmetric atmospheric circulation, are a manifestation of the irregularities on the earth’s surface, such as orography and land–sea thermal contrasts. There are substantial month-to-month variations in both the amplitude and spatial structure of the stationary waves. Understanding the forcing mechanisms and the seasonal cycle of the climatological stationary waves is of central importance in seasonal climate prediction. For example, regional climate anomalies, such as the 1988 drought and the 1993 floods, are associated with anomalous stationary waves over the Pacific and North American region, which are deviations from the climatological stationary waves (Liu et al. 1998). To obtain a better understanding of the possible causes for regional climate anomalies, it is necessary to determine the maintenance mechanisms of the climatological stationary waves. The objectives of this study are to document the month-to-month variations of the climatological stationary waves, and to determine the dominant forcing mechanisms for the maintenance and seasonal cycle of the stationary waves based on NCEP–NCAR reanalysis data and a linear stationary wave model.

As shown in many previous studies (e.g., Nigam et al. 1986, 1988; Chen and Trenberth 1988a,b; Valdes and Hoskins 1989; Ting 1994), the linear stationary wave...
model can be a powerful diagnostic tool for understanding the maintenance of stationary planetary waves. In the linear model, stationary waves are generated when the prescribed zonal-mean or zonally varying basic flow is subjected to various zonally asymmetric forcings. After assessing the validity of the linear model, further decomposition of the linear model response to those due to individual forcings can determine the dominant maintenance mechanisms of the climatological stationary waves.

The zonal-mean or zonally varying basic state and the zonally asymmetric forcings used in the linear model can be from general circulation model (GCM) output or from observations. The advantage of using GCM data is that the linear model and the forcings are dynamically consistent. Thus, the comparison of linear model solutions and the stationary waves produced by the GCM can provide a validation for the linear stationary wave theory. For example, Nigam et al. (1986, 1988) used the Geophysical Fluid Dynamics Laboratory (GFDL) GCM data and a linear model to understand the maintenance of the northern winter stationary waves with and without orography, and Ting (1994) used a similar model to diagnose the dominant forcing mechanisms for the northern summer GCM stationary waves. To what extent these results based on GCM data can be applied to the atmosphere remains to be determined. The advantage of using the observational data in stationary wave diagnosis is that the results are directly applicable to the real atmosphere. However, observational data are often subject to uncertainties due to instrumental errors and/or deficiencies in data processing techniques. Consequently, the forcing functions derived from observations lack the dynamical consistency required by the linear model. In particular, it is difficult to obtain an accurate three-dimensional diabatic heating rate, which was shown in many previous studies to be an important forcing in maintaining the climatological stationary waves (Valdes and Hoskins 1989; Ting 1994).

In recent years, the quality of the observational data has been greatly improved. The linear model diagnostics of stationary waves using observational data becomes possible. One such study was made by Valdes and Hoskins (1989), who used a 6-yr European Centre for Medium-Range Weather Forecasts analysis to examine the maintenance of northern winter climatological stationary waves. They obtained a reasonably good Northern Hemisphere (NH) simulation at the upper-tropospheric level. The model results over the Southern Hemisphere (SH) at the upper level and the global simulation at the lower level were rather poor, however. The poor simulation is partially due to the weak tropical heating used, which was a result of the deficiencies in the adiabatic initialization scheme (Valdes and Hoskins 1989). Given the uncertainties in both the studies based on GCM data and observational data, it is difficult to determine the nature of the differences among different studies. For example, Nigam et al. (1986, 1988) and Valdes and Hoskins (1989) reached different conclusions as to the relative importance of orography and diabatic heating for the maintenance of climatological northern winter stationary waves in NH extratropics. Nigam et al. (1986, 1988) showed that orography and diabatic heating are of equal importance, whereas Valdes and Hoskins (1989) concluded that diabatic heating is of predominant importance.

To solve the above problems, a high quality observational database is required. The ongoing National Centers for Environmental Prediction (NCEP) and the National Center for Atmospheric Research (NCAR) reanalysis project provides a suitable dataset for this purpose. By utilizing a frozen state-of-the-art data assimilation system and improved quality control, the NCEP–NCAR reanalysis provides a high quality dataset for the linear stationary wave simulations. In our study, only those variables strongly influenced by raw observations (horizontal wind components $U$ and $V$ and temperature $T$) and those that depend both on observations and model (pressure vertical velocity $\omega$ and surface pressure $P_s$) are used.

While the maintenance of northern winter and summertime stationary waves was studied in the past, there are no previous studies that address the seasonal cycle of the climatological stationary waves. There are no stationary wave simulations using observations for the northern summer season. With much improved and consistent 9-yr reanalysis data and a linear stationary wave model, this study aims at the investigation of the forcing mechanisms for the climatological stationary wave seasonal cycle. Based on the analysis of month-to-month stationary wave changes due to each forcing mechanism, the maintenance of the stationary wave seasonal cycle can be determined. In particular, the following questions will be addressed in this study.

1) To what extent is the linear stationary wave model qualitatively and quantitatively valid in simulating the climatological stationary waves and their seasonal cycle?
2) What are the main forcing mechanisms that account for the maintenance of the climatological stationary waves and their seasonal cycle?
3) What is the relative importance of diabatic heating and orography in NH extratropics in northern winter and summer?
4) What is the relative importance of local and remote thermal forcing in the maintenance of the stationary waves and their seasonal cycle?
5) What is the effect of basic state on the stationary wave seasonal cycle?

This paper is organized as follows. Following the introduction, a brief description of the NCEP–NCAR reanalysis data and the linear model is given in section 2. Sections 3 and 4 present the seasonal cycle of the derived three-dimensional diabatic heating field and the climatological stationary waves, respectively. Section 5 describes in detail the linear model simulations using zonal-mean basic state and the forcings, and the effects...
of the individual forcing mechanisms. Section 6 examines separately the effect of the seasonally varying zonal-mean basic state and that of the seasonally varying forcing fields on the seasonal cycle of the stationary waves. Finally, the summary and conclusions as well as a discussion of future research work are provided in section 7.

2. Methodology

a. Data

The data used in this study are taken from the NCEP–NCAR reanalysis project (Kalnay et al. 1996). In this project, a frozen state-of-the-art data assimilation system is used, which eliminates climate jumps due to the use of different data assimilation systems. Advanced quality control is applied to the data prior to the assimilation. The reanalysis data used in our study are in pressure coordinates. They have global coverage and high vertical resolution, with a total of 17 pressure levels. More details about the reanalysis data are given in Kalnay et al. (1996).

In this study, we use only those variables that strongly depend on raw observations (U, V, T) and those that depend on both model and observations (ω, PΓ). A total of 9 yr (1 January 1985–31 December 1993) of twicedaily data are used. The data are global (0°–360°E; 88.875°S–88.875°N), with a horizontal grid spacing of 2.5° long by 2.5° lat. The zonal and meridional wind (U, V) and temperature (T) are provided at 17 pressure levels, whereas vertical motion (ω) is available only at the lowest 12 pressure levels.

The transient momentum, heat, and vorticity fluxes are calculated explicitly from the twice-daily deviations from the corresponding monthly mean. Thus, all the calculated transient fluxes are of a submonthly timescale. The climatological mean of both the fluxes and the basic variables is obtained by averaging the monthly mean quantities from 1985 to 1993.

b. Linear baroclinic model

Similar to Ting (1994), the linear model employed in this study is a spectral, steady, linear baroclinic model in sigma (σ) coordinates. This model has rhomboidal wavenumber 30 truncation and 14 unevenly spaced vertical σ levels. The basic prognostic equations are those for vorticity, divergence, temperature, and log of surface pressure. The hydrostatic equation and mass continuity equation are also included to determine the geopotential height Φ and the sigma dot vertical velocity (σ). All equations are linearized about the reanalysis zonal-mean basic flow. More details about the model and the matrix inversion technique are referred to in Ting and Held (1990) and Ting (1994).

The dissipations employed in this linear model include Rayleigh friction (κ), Newtonian cooling (κΓ), and biharmonic diffusion (νΓ). They enter the basic equations as follows:

\[
\frac{\partial \zeta}{\partial t} = \ldots - \kappa \zeta - \nu \nabla^4 \zeta
\]

\[
\frac{\partial D}{\partial t} = \ldots - \kappa D - \nu \nabla^4 D
\]

\[
\frac{\partial T}{\partial t} = \ldots - (\kappa + \nu \nabla^4) \left( T - \frac{\partial [T]}{\partial \ln \sigma} \ln P \right)
\]

and the coefficients are chosen to be

\[
\kappa = \kappa_Γ = \begin{cases} 
(0.3 \text{ days})^{-1} & \sigma = 0.997 \\
(0.5 \text{ days})^{-1} & \sigma = 0.979 \\
(1.2 \text{ days})^{-1} & \sigma = 0.935 \\
(15 \text{ days})^{-1} & \text{elsewhere}
\end{cases}
\]

and \( \nu_Γ = 1 \times 10^{-7} \text{ m}^4 \text{ s}^{-1} \). Rayleigh friction and Newtonian cooling applied at the lowest three σ levels represent momentum and heat transfer within the boundary layer, while the dissipations in the free atmosphere represent partially the nonlinear effects. The highly scale-selective biharmonic diffusion is used to smooth out small-scale noises in the linear model solutions.

The basic state used in the linear model is taken from the 9-yr mean of the zonal-mean horizontal wind (U, V), temperature (T), and log of surface pressure (ln PΓ). The zonal-mean state varies from month to month. The latitude–month plots of the zonal-mean zonal wind at the jet stream level (σ = 0.257) and the surface (σ = 0.997) are shown in Figs. 1a and 1b, respectively. The mid-latitude westerly jet (Fig. 1a) in the winter hemisphere is much stronger than its counterpart in the summer hemisphere. The maximum zonal-mean zonal wind in the NH moves northward from April to July and southward from September to December. The equatorial easterlies follow the same north–south seasonal movement as that of the jet location and reach their maximum in northern summer months. At the surface (Fig. 1b), easterlies dominate throughout the year between 30°S and 30°N, with maximum reached in the winter season of both hemispheres. Between 30° and 60° in both hemispheres, westerlies prevail. The westerlies are much stronger in the SH than in the NH, due to the dominance of ocean surface and weak surface friction in the SH. Unless otherwise stated, the zonal-mean state of a particular month is used to simulate the stationary waves for that month.

The stationary wave forcings include orography, diabatic heating, transient forcing, and stationary linearity. The orographic forcing enters the hydrostatic equation as the lower boundary condition. The realistic orography employed in this study is taken from the NCEP–NCAR reanalysis data and is rhomboidally truncated at wavenumber 30. The seasonal variation of the orographic forcing is a result of the surface wind seasonal cycle as shown in Fig. 1b. The three-dimensional
The bar in (2) represents the monthly average and prime the deviation from the mean, and $TF_{\text{vort}}$, $TF_{\text{Div}}$, and $TF_{\text{Temp}}$ enter the right-hand side of the basic equations for vorticity, divergence, and temperature, respectively. The transient forcings are computed from the twice-daily data at pressure levels for each month and averaged over the 9-yr period. In addition to the above forcings (i.e., orography, diabatic heating, transient forcings), the non-linear stationary wave vorticity and heat fluxes, which have been neglected during the linearization, are computed based on the NCEP–NCAR reanalysis data. These terms are included in the right-hand side of the basic model equations as an additional forcing, that is, the stationary nonlinearity. This stationary nonlinear wave–
wave interaction forcing is used to crudely represent the total stationary nonlinear effect in the linear model. The stationary nonlinearity forcing is directly calculated based on (A17)–(A21) in the appendix of Ting (1994) using the monthly mean reanalysis data. More details about this forcing are referred to Ting (1994).

3. Derived diabatic heating

Due to the lack of direct measurement, various approaches have been used in the past to obtain the diabatic heating. Opsteegh and Verneker (1982) parameterized the heating in their model. Based on the First Global Atmospheric Research Program Global Experiment data, Johnson and Wei (1984) derived the diabatic heating using the vertically integrated mass continuity equation in isentropic coordinates. In addition, many studies (e.g., Youngblut and Sasamori 1980; Valdes and Hoskins 1989; Christy 1991) derived the diabatic heating rate as a residual in the thermodynamic equation. Due to the lack of mass and momentum balance in the data, the residually derived heating fields were subject to uncertainties. The diabatic heating fields obtained using different methods based on different data sources had considerable differences in both horizontal and vertical structures.

A recent study (Yanai and Tomita 1998) illustrated that the global diabatic heating can be accurately derived from the NCEP–NCAR reanalysis based on the budget of the thermodynamic equation in pressure coordinates. In this study, due to the availability of all data at pressure levels, the three-dimensional diabatic heating field is similarly derived from the thermodynamic equation in pressure coordinates using monthly mean data as follows:

$$\overline{\Omega} = \frac{\partial \overline{T}}{\partial t} + \nabla \cdot \nabla \overline{T} + \overline{\omega} \left( \frac{\partial \overline{T}}{\partial p} - \frac{R \overline{T}}{C_p \rho} \right) - T_{\text{temp}} F. \quad (3)$$

The monthly mean diabatic heating rate (\(\overline{\Omega}, \text{K s}^{-1}\)) is averaged over the 9-yr period to obtain the climatological mean heating for that month. Note that the residually derived diabatic heating contains all processes that are not resolved by Eq. (3). For example, damping and dissipations are lumped into \(\overline{\Omega}\). Here \(\overline{\Omega}\) also includes the temperature tendency due to the effect of analysis increment (Dee and da Silva 1998). One should also note that the NCEP–NCAR reanalysis produces the total diabatic heating as well. This heating is not used here because of the relatively poor simulation of the stationary waves in the linear model when the reanalysis heating is included. The poor simulation is mainly due to the analysis increment as discussed in Dee and da Silva (1998). The derived diabatic heating fields are interpolated to sigma surfaces for linear model calculation. It should be noted that diabatic heating can be derived directly in the sigma coordinates if reanalysis data are available on sigma surfaces. This would avoid errors introduced by interpolation from pressure to sigma surfaces. However, all reanalysis data are available to us on pressure surfaces, thus the pressure coordinates thermodynamic equation (3) is used.

Monthly mean heating fields are obtained for all 12 months. For brevity, we show only the heating distribution for January, April, July, and October to illustrate the seasonal cycle. The mass-weighted column-averaged diabatic heating fields for these four months are given in Fig. 2. During northern winter months (Fig. 2a), tropical diabatic heating is concentrated over the Indonesian and western Pacific regions, with additional tropical heating maxima over South America and Africa. The main tropical heating centers are located in the SH during this season. In the extratropical NH, the diabatic heating is characterized by two strong oceanic centers over the Pacific and Atlantic storm track regions as well as continental coolings over Eurasia and North America. During the boreal spring months (Fig. 2b), the three tropical heating maxima tend to move northward. In the meantime, the two storm track heatings in the NH extratropics weaken substantially, while the cooling over the Antarctic continent intensifies. During northern summer months (Fig. 2c), the striking heating feature is the combination of the monsoon heating over India and east Asia and the heating over the western Pacific. The tropical heating centers are largely in the NH at this time, except the South Pacific convergence zone (SPCZ). North of 40°N, there is very little heating present in the summer. In the SH, there are large coolings over the Antarctic. During the transitional months from northern summer to winter (Fig. 2d), the three tropical heating maxima move southward and back into the tropical SH. The two strong oceanic heatings in the storm track regions are reestablished. Compared to the heating in April, the heating in October is slightly stronger over both the Tropics and the NH extratropics. In the tropical eastern Pacific, strong coolings and a narrow band of the intertropical convergence zone persist throughout the whole seasonal cycle. The SPCZ is also discernible from all months in Fig. 2.

To verify the derived diabatic heating field, we show in Fig. 3 the distribution of the Climate Prediction Center (CPC) merged 9-yr (1985–93) mean precipitation (Xie and Arkin 1997) for the same four months. Since tropical heating is largely due to latent heat release, tropical precipitation is closely related to the tropical diabatic heating field. The comparison of Figs. 2 and 3 in the Tropics shows a qualitative agreement between the two fields throughout the whole seasonal cycle. The three tropical heating centers as well as their seasonal movements are similarly depicted by precipitation in Fig. 3. The largest discrepancy between the derived heating and the precipitation is that the tropical precipitation is narrower and more continuously distributed in the longitudinal direction. The qualitative agreement between derived heating and precipitation in the Tropics demonstrates the usefulness of the derived heating. In
Fig. 2. Column-averaged mass-weighted residually derived diabatic heating distribution for (a) Jan, (b) Apr, (c) Jul, and (d) Oct, from the NCEP-NCAR reanalysis data. Contour interval is 1 K day$^{-1}$ and the zero contour is omitted. Regions with heating over 0.5 K day$^{-1}$ are shaded.
Fig. 3. CPC merged analyzed precipitation distribution (1985-93) for (a) Jan, (b) Apr, (c) Jul, and (d) Oct. Contour interval is $2 \times 10^{-2}$ mm day$^{-1}$ and regions with precipitation above $6 \times 10^{-2}$ mm day$^{-1}$ are shaded.
the extratropics, the total diabatic heating comprises both the latent and sensible heat components, thus precipitation only partially represents the diabatic heating field. There is a good correspondence between heating and precipitation over the two oceanic storm track regions, however. The seasonal variation of the intensity and position of the storm track heatings is well reflected in precipitation as well. The maximum amplitude of the storm track precipitation is only about half of that of the tropical precipitation, while the storm track heating is of the same magnitude as the tropical heating. This discrepancy can be attributed to the contribution of sensible heat to the total storm track heating in Fig. 2. By comparing the anomalous outgoing longwave radiation over the United States in the summers of 1988 and 1993 with the derived diabatic heating anomalies, Liu et al. (1998) found good agreement between the two fields based on the same data. This lends further support for the reliability of the derived diabatic heating in the mid-latitudes in this study. The derived heating in Fig. 2 also compares well with that in Yanai and Tomita (1998).

To illustrate the vertical structure of the derived diabatic heating, we show the zonal-mean heating distribution for the four months in Fig. 4. Throughout the four seasons, the tropical heating maximum in Fig. 4 is consistently located between 450 and 500 mb, whereas the midlatitudinal heating maximum is at approximately 850 mb. Deep cooling is present in the subtropics and high latitudes, with double maxima at 250 and 700 mb, respectively. The large-amplitude cooling over the south polar region during NH spring, summer, and fall may be due to the extrapolation of the wind and temperature fields beneath the Antarctic. In April, there is a clear split in tropical heating centers north and south of the equator. This feature is present in January to a lesser extent. The seasonal movement of the tropical heating centers as well as the seasonal intensity variation of the extratropical heatings are also well depicted in Fig. 4.

Compared to the residually calculated December–January–February averaged diabatic heating in Valdes and Hoskins (1989) (Fig. 2 from Valdes and Hoskins 1989), the heating shown in Figs. 2a and 4a is twice as strong as theirs in the Tropics, whereas the extratropical heating in the two storm track regions is approximately of the same magnitude. In addition, the derived continental cooling as well as the cooling over the tropical eastern Pacific in Fig. 2a are much more intense than those in Valdes and Hoskins (1989). Valdes and Hoskins (1989) recognized the weak tropical heating in their study as a result of the lack of a strong meridional circulation in the early analysis.

The seasonal variation of the diabatic heating field is further examined using the zonal-mean diabatic heating amplitude. The diabatic heating amplitude is defined as the square of the mass-weighted column-averaged diabatic heating rate ($\bar{Q}$):

$$\left( \frac{\int_{100\text{mb}}^{850\text{mb}} \bar{Q} \, dp}{\int_{100\text{mb}}^{100\text{mb}} dp} \right)^2.$$  

Figure 5 shows the seasonal variation of the zonally averaged diabatic heating amplitude as a function of latitude. In the northern winter, there are distinct tropical and extratropical heating regimes. The tropical regime is located at about 15°S and reaches its maximum in February, while the extratropical maximum is attained at about 40°N in January. The extratropical maximum is associated with the oceanic heatings as well as the continental coolings in the NH. During northern spring, the tropical heating center in Fig. 5 jumps abruptly to about 10°N, where the heating center remains throughout the northern summer. The extratropical heating is of negligible amplitude from April to September. There is only a single tropical heating regime in the northern summer. During northern autumn, the winter heating pattern reestablishes itself. In the SH extratropics, the heating amplitude is dominated by Antarctic cooling, particularly from March to October. In summary, the dominant seasonal variation of diabatic heating is the north–south movement of the tropical heating and the intensity variation of the extratropical heating and cooling.

4. Climatological stationary wave seasonal cycle

Since streamfunction emphasizes stationary waves in the Tropics and subtropics as well as at high latitudes, the zonally asymmetric streamfunction field is used to represent stationary waves in this study. The climatological mean streamfunction fields are obtained from the horizontal winds. The climatological stationary waves are shown at two $\sigma$ levels, that is, $\sigma = 0.866$ and $\sigma = 0.257$. They are referred to as the lower- and upper-tropospheric levels, respectively.

The climatological eddy streamfunction fields at the lower-tropospheric level for January, April, July, and October are shown in Fig. 6. The distinct NH winter stationary wave features are the Aleutian low, the high over the west coast of North America, the Icelandic low, as well as the Azores High. The above highs and lows are centered at around 45°N. From January to April, the stationary wave pattern changes dramatically. Almost all the NH winter features disappear and are replaced by the two oceanic highs over the Pacific and Atlantic Oceans and a weak low over Tibet. The subtropical highs over the Pacific and Atlantic as well as the low over Tibet intensify during NH summer and reach their maximum strength at 30°N in July. An anticyclone also persists over the Indian Ocean from May to September, which is part of the NH summer monsoonal flow over that region. From northern summer to fall, all the northern winter stationary wave features are in the process of being reestablished. The amplitude of the NH stationary waves reaches its minimum in October. Com-
Fig. 4. Latitude–pressure distribution of zonally averaged diabatic heating for (a) Jan, (b) Apr, (c) Jul, and (d) Oct. Contour interval is 0.5 K day$^{-1}$, and regions with heating over 0.5 K day$^{-1}$ are shaded.
compared to the NH stationary waves, the stationary waves in the SH at the lower level are weaker in general and have little seasonal variation in center locations.

The eddy streamfunction fields at the upper-tropospheric level are shown in Fig. 7. The stationary waves at the upper level are more pronounced in northern winter than any of the other seasons. In January (Fig. 7a), wavenumbers 2 and 3 dominate in the NH high latitudes, while wavenumbers 1 and 2 dominate over the Tropics and subtropics. The characteristic northern winter stationary wave features over the Tropics and subtropics include the strong and extensive anticyclone over the western tropical Pacific and the two tropical oceanic troughs over the eastern Pacific and Atlantic. In the NH midlatitudes, the dominant northern winter stationary wave features are the east Asian low, a high over the west coast of North America, a low over Hudson Bay, and a high over the North Atlantic. All of these extra-tropical features have their counterparts at the lower level (Fig. 6), indicating an equivalent barotropic vertical structure. From northern winter to spring (Fig. 7b), all the wintertime features are present but with weaker amplitude. During northern summer (Fig. 7c), NH stationary waves are dominated by the Tibetan anticyclone, two oceanic troughs in the Pacific and Atlantic, as well as an isolated anticyclone over central North America.

Over North America and North Atlantic regions, the high east of the Rocky Mountains, the low over northeast Canada, and the ridge farther downstream over the North Atlantic form a "great circle" wave train. The major stationary wave features in Fig. 7c at 30°N have their corresponding low-level counterparts in Fig. 6c, but with opposite polarity, suggesting the baroclinic vertical structure of summertime stationary waves. In October (Fig. 7d), the northern winter stationary wave features in the NH extratropics quickly reestablish, while the tropical and subtropical centers resemble the summer features. Similar to the lower level, the upper-level stationary waves in SH have much less seasonal variation in both amplitude and spatial structure than those in the NH.

To illustrate the full seasonal cycle of the stationary waves, we show in Fig. 8 the zonally averaged root-mean-square (rms) eddy streamfunction amplitude

$$\text{rms}(\psi^*) = \sqrt{\frac{1}{2\pi} \int_0^{2\pi} \psi^{*2}a \cos \theta d\lambda},$$

where $\psi^*$ is the stationary wave streamfunction, and $\lambda$ and $\theta$ are longitude and latitude, respectively, as a function of latitude and month at the lower (Fig. 8a) and upper (Fig. 8b) levels. In the NH extratropics, the low-level stationary waves in northern winter have relatively weak strength, with the centers located at around 45°N. The maximum stationary wave amplitude is reached at around 30°N in July, whereas the weakest stationary wave amplitude is found in October and November. In the SH, the stationary wave amplitude has a similar seasonal cycle to that in the NH, but the amplitude fluctuation is much weaker. At the upper level (Fig. 8b), the seasonal cycle of the stationary waves shows distinct differences from the lower level. There are two distinct maxima in northern winter, one at 50°N and another at 20°N. The strongest stationary wave amplitude is obtained in the winter subtropics. The northern summer stationary waves are weaker than those in the northern winter, with a single maximum at about 30°N. The weakest stationary wave amplitude is obtained during the two transitional seasons. It is interesting to note the striking similarity between the seasonal variation of the diabatic heating amplitude (Fig. 5) and that of the upper-level stationary wave strength (Fig. 8b). This implies the possible importance of diabatic heating in forcing the stationary wave seasonal cycle in the upper troposphere. Similar to the lower level, the seasonal variation of stationary waves in the SH at the upper level is much weaker. The possible role of diabatic heating, orography, and the transient heat and momentum flux convergences in forcing the climatological stationary wave seasonal cycle will be examined more quantitatively using the linear model in the next section.
Fig. 6. Climatological mean zonally asymmetric streamfunction at $\sigma = 0.866$ for (a) Jan, (b) Apr, (c) Jul, and (d) Oct. Contour interval is $3 \times 10^5$ m$^2$ s$^{-1}$ and regions with negative values are shaded.
Fig. 7. Same as Fig. 6 but for $\sigma = 0.257$. 
Fig. 8. Zonal rms streamfunction amplitude of the climatological stationary waves at (a) $s = 0.866$, (b) $s = 0.257$. Contour interval is $1 \times 10^6$ m$^2$ s$^{-1}$. Areas with value above $5 \times 10^6$ m$^2$ s$^{-1}$ in (a) and $8 \times 10^6$ m$^2$ s$^{-1}$ in (b) are shaded.
5. Linear model simulation and diagnostics of the stationary waves and their seasonal cycle

a. Linear model response to total stationary wave forcing

If the linear stationary wave theory is valid, climatological stationary waves can be simulated in the linear model by prescribing the zonally symmetric basic state and subjecting it to all the stationary wave forcings. As a first-order modification to the pure linear theory, one can add to the linear model an additional forcing term that represents the stationary nonlinear effect (Valdes and Hoskins 1989; Ting 1994). The linear model solutions, when forced by the sum of global orography, diabatic heating, the transient fluxes, and stationary nonlinearity, are shown in Figs. 9 and 10 for January, April, July, and October at the lower- and upper-tropospheric levels, respectively. At both the lower and upper levels, there is a good correspondence between the linear model results (Figs. 9 and 10) and the reanalysis climatological stationary waves (Figs. 6 and 7). The largest differences are that the tropical centers at the lower level for all months and the NH centers at the upper level for northern spring, summer, and fall are much too strong in the linear model. The discrepancy may be due to the inaccuracies in the derived diabatic heating fields, since diabatic heating is the most dominant forcing mechanism in the Tropics, as will be shown later. Large discrepancies are also present over the Antarctic and Andes from April to October, due to both the orographic forcing and the unrealistic cooling there.

To further determine how well the linear model simulates the spatial structure of the reanalysis climatological stationary waves, the area-weighted spatial pattern correlation between the climatological stationary waves and the model simulation is calculated. The pattern correlation $r$ is defined as

$$
r = \frac{\int_A \int (\psi^{\text{reanalysis}} \psi^{\text{model}} \cos \theta) \text{d} \lambda \text{d} \theta}{\left[ \int_A \int (\psi^{\text{reanalysis}} \cos \theta) \text{d} \lambda \text{d} \theta \right] \left[ \int_A \int (\psi^{\text{model}} \cos \theta) \text{d} \lambda \text{d} \theta \right]^{\frac{1}{2}}},
$$

where $\psi^{\text{reanalysis}}$ and $\psi^{\text{model}}$ denote the reanalysis- and linear model–simulated streamfunction perturbation, respectively, and “$A$” indicates the area over which the spatial pattern correlation is calculated. Figure 11 shows the spatial pattern correlation as a function of month and vertical level for stationary waves over (a) the global domain, (b) the NH, and (c) the SH. In all three diagrams, the largest pattern correlations are found in northern winter and summer months at the upper-tropospheric levels. The stationary waves at the lower- and middle-tropospheric levels are not as well simulated as those at the upper levels. One exception is the NH summer (June, July, and August), for which the linear model simulation below $\sigma = 0.6$ is almost as good as that at the upper levels. The pattern correlation reaches its minimum in April and May for the whole troposphere. By comparing the three diagrams, it is evident that the stationary waves are better simulated in the NH than in the SH. When the stationary nonlinearity is not included in the total forcing, the spatial pattern correlation drops slightly everywhere in the domain (not shown), indicating the importance of the nonlinearity in modifying the stationary wave patterns. However, the general characteristics of the pattern correlation remain the same as those in Fig. 11. The modest improvement by including the stationary nonlinearity effect indicates that the linear stationary wave theory is qualitatively valid, at least in the upper troposphere. The lower pattern correlation in the middle and lower troposphere is partially contributed by the unrealistic pattern in the Tropics in Fig. 9 compared to that in Fig. 6. Both inaccuracies in data and linear model formulation may contribute to the discrepancy. It is interesting to note that while the highest correlation in the SH is not as high as that in the NH, the pattern correlation in the middle troposphere is higher in the SH than that in the NH. Given that data quality in the SH is generally lower than that in the NH, it is possible that the discrepancy between linear model results and reanalysis climatology in the lower troposphere is more related to the linear model formulation, such as the damping used.

Figure 12 illustrates the seasonal variation of the zonally averaged rms stationary wave amplitude due to the total forcing at the lower- (Fig. 12a) and upper- (Fig. 12b) tropospheric levels as a function of latitude and month. Compared to the corresponding reanalysis (Fig. 8), the linear model provides a reasonable simulation of the climatological stationary wave seasonal cycle. The main deficiencies of the linear model simulation are the amplitude overestimation in the Tropics at the lower level for all months and in both the Tropics and extratropics at the upper level from October to April, and a shift in timing of the maximum center at the upper level. The linear model fails to reproduce the northern summer stationary wave maximum at 30°N in the upper troposphere. The linear model also produces an unrealistically large response in Antarctic at both lower and upper levels.

Overall, the linear model provides a reasonably good simulation of the stationary wave seasonal cycle. This provides a basis for the further decomposition of linear model simulated stationary waves into different components forced by each individual forcing mechanism, which will be presented in the following sections.

b. Linear model responses to individual forcings

The linear model response to each of the four stationary wave forcings, that is, orography, diabatic heating, transients, and stationary nonlinearity, is calculated for each month. For brevity, only the upper-tropospheric
Fig. 9. Zonally asymmetric streamfunction response at $\theta = 0.866$ to the total forcing (diabatic heating, orography, stationary nonlinearity, and transient forcing) for (a) Jan, (b) Apr, (c) Jul, and (d) Oct. Contour interval is $3 \times 10^6$ m$^2$ s$^{-1}$, and regions with negative values are shaded.
Fig. 10. Same as Fig. 9 but for $\sigma = 0.257$. 
responses to different forcings in January and July will be presented. The complete seasonal cycle of the stationary waves due to each individual forcing will be discussed in terms of the zonal rms stationary wave amplitude.

Figure 13 shows the upper-level streamfunction responses to diabatic heating (a), orography (b), stationary nonlinearity (c), and transient eddy flux convergences (d) in January. By comparing Fig. 13 to the model response to total January forcing in Fig. 10a, diabatic heating is evidently the major contributor to the total response over the global domain. It is particularly true for stationary waves over the Tropics and the SH. Almost all the January highs and lows in Fig. 10a have their counterparts centered at the same locations in the model response to diabatic heating only (Fig. 13a). The contribution by stationary nonlinearity is significant in the NH (Fig. 13c). For example, the anticyclone centered over the west coast of North America is largely maintained by stationary nonlinearity, while diabatic heating only plays a secondary role. The role of stationary nonlinearity in maintaining the anticyclone over northwestern North America was also found in Valdes and Hoskins (1989). As indicated in Fig. 13c, the stationary nonlinearity helps to extend the western Pacific anticyclone eastward. Also, there is a tendency for stationary nonlinearity to produce an anticyclone both over and southeast of the major mountain ranges in the NH, and a cyclonic center northeast of the mountains. The close connection between orographic features and the response to stationary nonlinearity suggests the possible effect of the nonlinear orographic forcing. The nonlinear stationary waves forced by an idealized orography were discussed in Ringler and Cook (1997). In this study, it is not possible to determine the exact nature of the nonlinear forcing using a linear model. The linear effect of orography (Fig. 13b) is relatively small compared to the effects of heating and nonlinearity. The most distinctive feature of the linear model response to orography is the two wave trains emanating from the mountain ranges in the NH, which has been illustrated in many previous studies (e.g., Grose and Hoskins 1979; Hoskins and Karoly 1981). In the NH, distinct troughs lie east of the Tibetan Plateau and the Rockies, and a weak wave train propagates to the southeast and dissipates as it approaches the tropical zonal-mean easterlies. The orographic response is too weak to account for the climatological stationary wave features. The linear model response to transient forcing is relatively weak, but this forcing is found to dissipate the total model response from eastern Asia to the eastern Pacific in the Tropics. In the SH, the effects of orography, stationary nonlinearity, and transient forcing are rather weak in January.

For brevity, the low-level responses to different forcing components in January are not shown. But it is interesting to note that the NH extratropical response to stationary nonlinearity forcing at the lower level (Fig. 14) reproduces almost all features of the corresponding response to total forcing (Fig. 9a) in January. The predominant importance of the stationary nonlinearity effect in the NH extratropics at the lower level during northern winter is worthy of further investigation.

Figure 15 illustrates the model responses to the four forcings in July at the upper-tropospheric level. Similar to January, the relative importance of the forcings can be quantitatively determined by comparing the model response to each forcing (Fig. 15) and that to the total...
Fig. 12. Zonal rms streamfunction amplitude of stationary wave response to total forcing at (a) \( \sigma = 0.866 \) and (b) \( \sigma = 0.257 \). Contour interval is \( 1 \times 10^{6} \) m$^2$s$^{-1}$. Areas with value above

(a) \( 5 \times 10^{6} \) m$^2$s$^{-1}$ and (b) \( 8 \times 10^{6} \) m$^2$s$^{-1}$ are shaded.
Fig. 13: Zonally asymmetric streamfunction at $\sigma = 0.257$ forced by (a) global diabatic heating, (b) orography, (c) stationary nonlinearity, and (d) transient forcing for January in the linear model. Contour interval is $3 \times 10^{-6} \text{ m}^2 \text{s}^{-1}$ and regions with negative values are shaded.
forcing (Fig. 10c). In the Tropics, diabatic heating plays the dominant role. The magnitudes and locations of the tropical centers in the model response to diabatic heating are comparable to those in the model response to total forcing. In the NH extratropics, the streamfunction response to diabatic heating is much weaker than that in the Tropics, but still accounts for a significant fraction of the response to total forcing in Fig. 10c. The effect of stationary nonlinearity is comparable to that of heating during summer in the NH extratropics. It is the only factor that accounts for the anticyclone over central North America. This effect of the stationary nonlinearity over the United States in northern summer is consistent with the results in Ting (1994) using the GFDL GCM data. The linear model response to orography in the NH in July is negligible. The linear model generates larger-amplitude response to orography over the South Pole region, which does not have a counterpart in the reanalysis. The transient forcing has a negligible effect in July (Fig. 15d) over the whole domain.

To diagnose the relative importance of these four forcings in contributing to the seasonal variation of the total stationary wave amplitude, we show the month-to-month variation of the zonal rms stationary wave amplitude due to each forcing at the upper level in Fig. 16. By comparing the four diagrams in Fig. 16 with Fig. 12b, it is clear that in the Tropics, diabatic heating has the predominant contribution to the seasonal cycle of the stationary wave amplitude. By contrast, the effects of the other three forcings are negligible in the Tropics. In the NH extratropics, diabatic heating accounts for about half of the seasonal cycle of stationary wave amplitude due to the total forcing (Fig. 12b), whereas orography contributes less than one-third of the total amplitude during northern winter and even less for the other three seasons. The stationary nonlinear effect is significant, but there is less organized seasonal variation in amplitude. Transient forcing tends to contribute more at the high latitudes in northern winter than at the midlatitudes where the reanalysis climatological maximum is located. Similar conclusions as to the relative importance of the individual forcings can be drawn at the lower level (not shown).

c. Linear model response to tropical and extratropical heating

Due to the predominant contribution from diabatic heating, one important question to address is the relative importance of local and remote heating on the maintenance of stationary waves in both the Tropics and the extratropics. It is well known that tropical heating can exert an important forcing on the extratropical atmosphere (Hoskins and Karoly 1981; Simmons 1982). It is thus interesting to examine the remote effect of tropical heating on the extratropical stationary wave seasonal cycle in the reanalysis data. Based on the spatial distribution of diabatic heating (Fig. 2), we define tropical heating to be the heating within the belt from 35°S to 25°N and NH extratropical heating to be the heating within the belt from 25°N to the North Pole. The relative importance of the tropical and extratropical heating is determined by performing linear model simulations forced by only the tropical or NH extratropical heating. The results are shown in Fig. 17, which illustrates the zonal rms stationary wave amplitude at the upper and lower levels as a function of latitude and month. Figure 17 demonstrates
Fig. 15. Same as Fig. 13 but for the streamfunction response in Jul.
clearly the dominant effect of local heating in forcing the local stationary wave seasonal cycle. The remote effect of diabatic heating, on the other hand, is much smaller compared to the local effect of local heating.

**d. Effect of dissipations**

The dissipations are shown to be effective in removing the critical line singularity over most of the global regions. However, when Rayleigh friction and Newtonian cooling are sufficiently large, model solutions are insensitive to the choices. Larger dissipation coefficients only slightly decrease the stationary wave amplitude, and the stationary wave patterns remain unchanged. On the other hand, compared with $1 \times 10^{16}$ m$^4$ s$^{-1}$, the stronger biharmonic coefficient $1 \times 10^{17}$ m$^4$ s$^{-1}$ is shown to make greater improvements over the SH high latitudes. The amplitudes of the centers close to the South
Fig. 17. Zonal rms streamfunction amplitude due to NH extratropical diabatic heating (25°–88.875°N) at (a) $\sigma = 0.257$, (b) $\sigma = 0.866$, and tropical diabatic heating (35°–25°N) at (c) $\sigma = 0.257$, (d) $\sigma = 0.866$. Contour interval is $1 \times 10^6$ m$^2$ s$^{-1}$. Areas with value above $5 \times 10^6$ m$^2$ s$^{-1}$ are shaded.

Pole are reduced significantly, whereas those for the rest of the global regions are much less affected.

6. The effect of seasonally varying zonal-mean basic state and the forcing

As shown in Fig. 1, the zonal-mean flow has substantial variations from one month to another. Ting et al. (1996) found that a significant fraction of the interannual atmospheric variability in the NH extratropics can be explained by the zonal-mean flow fluctuations in winter. The seasonal fluctuation of the zonal-mean flow is much stronger than the interannual fluctuations shown in Ting et al. (1996), it is thus important to examine the contribution of the seasonal fluctuations of zonal-mean flow to the seasonal cycle of the stationary waves. The results presented in the previous section combine the effect of zonal-mean flow fluctuations and
that of the seasonal changes of the forcing fields, in particular, the diabatic heating. To separate the effect of the seasonally varying zonal-mean basic state from that of the seasonally varying forcing fields, we perform two simple calculations using the linear model. In the first calculation, the total forcing is prescribed in the linear model at the annual-mean value while the zonal-mean basic state is allowed to vary from month to month. In the second calculation, the zonal mean flow is fixed at the annual-mean state while the total forcing varies from month to month. The stationary wave seasonal cycle is examined using the zonal rms amplitude of the streamfunction at upper level in Fig. 18. Without the seasonal variation of the stationary wave forcings (Fig. 18a), the stationary wave amplitude fluctuations at the upper level are well reproduced in the Tropics between the equator and 30°N, compared to Fig. 12b. But the seasonal fluctuation of the extratropical stationary wave amplitude is much weaker in Fig. 18a than that in Fig. 12b. On the other hand, the effect of the seasonally varying total forcing alone (Fig. 18b) is able to largely explain the zonal-mean stationary wave amplitude fluctuations over the NH extratropics. Figure 18b also reproduces well the amplitude maximum in northern summer, although the center is shifted farther south compared to that in Fig. 8b.

It is also interesting to examine the seasonal change in stationary wave spatial structure due to the effect of zonal-mean flow and that of total forcing. Figure 19 illustrates the spatial pattern correlation between linear model results and reanalysis in the two cases for the global (top), NH (middle), and SH (bottom) domain. From September to May, the spatial pattern correlation in both cases is only slightly lower than that in Fig. 11 throughout the troposphere. Thus the stationary wave seasonal transition from northern winter to spring, and from northern autumn to winter, can be well reproduced by varying only the zonal-mean basic state or the stationary wave forcings. For the northern summer months, however, the correlation is much lower in the case of varying only the zonal-mean flow (left panels) at both upper and lower levels, compared to the case of varying the total forcing field. The results in Fig. 19 clearly illustrate that the change in monsoon heating is essential for getting the northern summer stationary waves.

In summary, the seasonal variation of the zonal-mean zonal flow is very important in accounting for the stationary wave amplitude fluctuations in the Tropics from September to May. The amplitude variation in the extratropics, however, cannot be explained by the seasonal variation of the zonal-mean zonal flow alone. It is largely due to the seasonally varying forcing fields. Furthermore, the transition to the northern summer stationary wave regime cannot be explained without the accompanying changes in the heating field.

7. Summary and discussions

The climatological stationary wave seasonal cycle is examined in this study, with the emphasis laid on the maintenance mechanisms of the stationary waves throughout the seasonal cycle. By using the 9-yr (1985–93) NCEP–NCAR reanalysis product (Kalnay et al. 1996), the three-dimensional climatological diabatic heating field is first derived as a residual in the pressure coordinate thermodynamic equation. The derived heating depicts well the seasonal cycle of the global heating and provides a reasonable comparison to the corresponding precipitation field. The climatological stationary waves, in the form of the zonally asymmetric streamfunction, are also calculated based on the NCEP–NCAR reanalysis data. The climatological stationary wave fields show significant month-to-month fluctuations, in both spatial pattern and amplitude. During the northern winter months (October–April), stationary waves show tropical and extratropical regimes, centered at 20° and 50°N, respectively, while during the northern summer months (May–September) stationary waves are generally centered at about 30°N. There is much less seasonal variation in both stationary wave pattern and amplitude in the Southern Hemisphere.

To determine the maintenance mechanisms of the climatological stationary wave seasonal cycle in the NCEP–NCAR reanalysis, a spectral, steady, σ-coordinate, linear primitive equation model is used. When the model is linearized about the monthly mean zonally averaged basic state and subjected to certain dissipations and the zonally asymmetric stationary wave forcings, including diabatic heating, orographic uplifting, transient heat and vorticity flux convergences, and stationary nonlinearity, the model solutions for all 12 months are shown to have reasonably good agreement with the reanalysis in both stationary wave pattern and amplitude. This is especially true in the upper troposphere for all months as well as in the lower troposphere during northern summer. The general agreement between the reanalysis climatological stationary waves and the linear model simulation implies the reliability of the residually derived three-dimensional diabatic heating field.

Since the model with total forcing can reproduce the climatological stationary waves for each month reasonably well, the maintenance of the stationary waves throughout the seasonal cycle is determined by decomposing the total model response to those due to individual forcings. It is found that, in the Tropics, diabatic heating has the predominant contribution to the maintenance of the stationary waves throughout the seasonal cycle. In the extratropics, diabatic heating accounts for more than 50% of the stationary wave amplitude in northern winter months. During northern winter, most of the NH extratropical stationary wave features are accounted for by the stationary nonlinearity effect. The effect of stationary nonlinearity is particularly important over extratropical North America at the upper level and over the whole extratropics at the lower level. Consistent with the findings in Ting (1994), the effect of stationary nonlinearity is the dominant mechanism to maintain the anticyclone over the United States during northern sum-
Fig. 18. Zonal rms streamfunction amplitude at $\sigma = 0.257$, obtained from the linear model (a) with seasonally varying zonal-mean basic state and annual mean total forcing, and (b) with seasonally varying total forcing and annual-mean basic state. Contour interval is $1 \times 10^6$ m$^2$ s$^{-1}$ and areas with value above $3 \times 10^6$ m$^2$ s$^{-1}$ are shaded.
FIG. 19. Spatial pattern correlation between reanalysis climatology and linear model simulation with seasonally varying zonal-mean basic state and annual-mean total forcing over (a) the global domain, (b) the NH domain, and (c) the SH domain, and between reanalysis climatology and linear model simulation with seasonally varying total forcing and annual-mean basic state over (d) the global domain, (e) the NH domain, and (f) the SH domain. Contour interval is 0.1 and areas with value above 0.8 and below 0.6 are dark and lightly shaded, respectively.

mer. There are indications that the effect of stationary nonlinearity is closely related to the orographic features in the NH. Although stationary nonlinearity has significant amplitude throughout the seasonal cycle in both the Tropics and extratropics, it does not show the same amplitude fluctuation as a function of month as the climatological stationary waves. In the NH extratropics, the linear effect of orography explains only about 20% of the total stationary wave amplitude during northern winter and even less in the other three seasons. Orography is unimportant in the Tropics throughout the whole seasonal cycle. The effect of transients is small overall. The relative importance of local heating and remote heating in contributing to the local stationary wave seasonal cycle due to global diabatic heating is also examined in this study. It is found that local heating dominates the local model response to global heating in both the Tropics and extratropics.

The relative contribution of the effect of varying only the zonal-mean flow and that of varying only the
total forcing fields to the stationary wave seasonal cycle is assessed. The effect of the zonal-mean basic state seasonal cycle contributes significantly toward the amplitude fluctuations in the Tropics. It also contributes significantly to the seasonal transition of the stationary wave spatial structure from northern winter to spring and from northern autumn to winter. The effect of the total stationary wave forcing seasonal cycle, on the other hand, dominates the stationary amplitude fluctuation in the extratropics and the northern summer season. The establishment of the summer monsoon heating is essential for capturing the northern summer stationary wave pattern at both the upper and lower levels.

This study stresses the important role played by the global diabatic heating in maintaining the climatological stationary waves and their seasonal cycle. Our results confirm the findings in Valdes and Hoskins (1989) that the linear effect of the orography is secondary in the maintenance of stationary waves in NH winter extratropics. Through the use of a state-of-the-art re-analysis product and a complete examination of the whole seasonal cycle, this paper is able to address more quantitatively the role of each forcing mechanism at different times of the year and at both the upper and lower levels. Several problems remain unanswered in this study. First, the accuracy of the diabatic heating as derived from the residual of the thermodynamic equation is still uncertain. This contributes to the over-estimation of the stationary wave amplitude in most of the months. Second, the predominant importance of the stationary nonlinearity effect in northern winter extratropics raises the question of the ultimate cause of the nonlinearity. This problem cannot be answered in a linear framework. Extension to a fully nonlinear model is necessary to address the nature of the stationary nonlinearity. Third, the cause for the weak global pattern correlation during transitional months (e.g., April, May) is not determined. This problem can be addressed in a similar study but with GCM data instead of the observed data, since GCM data are dynamically consistent and thus one expects that lower pattern correlation reflects certain model deficiencies. In the future, the seasonal cycle of the GFDL GCM (Alexander and James 1995) will be analyzed using the same linear model. The maintenance of the GCM stationary wave seasonal cycle will be examined. Moreover, the comparison between the GCM and observations can reveal the deficiencies and possible improvements of the GCM simulation. The predominant importance of the diabatic heating in accounting for the stationary wave seasonal cycle also raises the issue of the importance of an accurate representation of the heating field in GCMs on the simulation of stationary waves and their seasonal cycle.

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REFERENCES


