Transient Cyclone-Scale Vorticity Forcing of Blocking Highs

WERNER METZ

Meteorological Institute, University of München, FRG

(Manuscript received 10 July 1985, in final form 5 February 1986)

ABSTRACT

The forcing of Northern Hemisphere blocking highs by the transient vorticity transfer from the cyclone-scale eddies into the planetary flow is investigated. Thereby, a barotropic prediction model for the planetary modes is numerically integrated. The time-dependent cyclone-scale vorticity forcing is evaluated from the observations.

The blocking activity in the forced model flow is examined by means of an objective analysis scheme. It is found that the model flow exhibits frequent blocking with some regional preference for the Atlantic and Pacific oceans. The mean structure of the model blocks compares well with the observations, particularly in the case of Atlantic blocking. However, the mean amplitude of the model blocks is only about two-thirds of those observed. Furthermore, it is found that the time-mean part of the forcing does not support the blocking pattern, so that the model blocking activity is due only to the transient forcing.

The significance of the cyclone-scale vorticity forcing for the occurrence of observed blocks is established in terms of model flow pattern averaged over days of actually observed blocking. In the case of Atlantic blocks, the model reproduces the correct mean structure of the observed flow with about one-third of the observed intensity. On the other hand, this mechanism fails in the Pacific. It is believed that this shortcoming is partly due to poor data resolution in the Pacific.

1. Introduction

Blocking anticyclones hinder the eastward propagation of synoptic-scale atmospheric disturbances. Thus, one may think that the large-scale blocking pattern acts as a steering mechanism for the small-scale eddies. On the other hand, the deflection of the traveling highs and lows by a block may lead to systematic anomalies of the transient eddy momentum and heat fluxes. In turn, these anomalies may feed back upon the planetary-scale flow. Accordingly, Green (1977) suggested that the time-mean transient-eddy vorticity fluxes may be important for the maintenance of a blocking split-flow pattern. More recently, Illari and Marshall (1983) computed the limited-area potential vorticity budget for one month of strong European blocking. They found that the time mean eddy fluxes helped to maintain the block. Studies of observed energy spectra by Hansen and Chen (1982) and Hansen and Sutera (1984) suggest that during blocking the planetary waves (wavenumber $m < 6$) systematically receive kinetic energy by the nonlinear interaction with all other waves. At the same time the cyclone-scale waves appear to lose kinetic energy via the same mechanism. Similar findings have also been made in a January climate run of a general circulation model (Fischer, 1984).

There have also been attempts to model the impact of the transient synoptic-scale eddies on the large-scale flow. Kalnay-Rivas and Merkine (1981) forced a barotropic channel flow by the release of local periodic vorticity pulses upstream of a narrow mountain barrier. Interestingly enough, they found that the transient forcing was able to produce a steady blocking configuration in the standing wave pattern downstream of the mountain. More recently, Shutts (1983) imposed periodic small-scale vorticity forcing upstream of a barotropic split-flow basic state. He showed that via an “eddy-straining” mechanism the small-scale eddies may enforce a quasi-steady dipole pattern, even in the absence of mountains, provided the zonal mean wind is weak enough.

Recently, Egger and Schilling (1983, hereafter referred to as ES) proposed a hypothesis concerning the low-frequency variability of the atmosphere. They consider a prognostic equation for the planetary scales of barotropic motion, forced by the transient eddy vorticity fluxes between the synoptic and the planetary scales. In response to these fluxes their equivalent-barotropic model was able to reproduce a large fraction of the observed low-frequency variance. In a subsequent paper (Egger and Schilling 1984), they found that, in particular, the slow modes of the planetary flow compare well with the theory of stochastically forced flows.

In view of the fact that blocking is related to the slow planetary modes of the atmospheric flow, we will investigate in the present paper whether transient cyclone-scale forcing of the planetary flow can induce and sustain blocking episodes. To elaborate this hy-
hypothesis we adopt the prognostic equation for the planetary modes from ES. This equation is integrated numerically, with the cyclone-scale forcing prescribed as an empirical function of time. This forcing is computed from observed wintertime data. The model is restricted to the Northern Hemisphere.

Let us emphasize that in the present study we want to investigate only one possible blocking mechanism, namely the impact of the cyclone-scale eddies on the planetary flow. Therefore, other possible blocking mechanisms, e.g., the direct effect of stationary (orographic or thermal) forcing on the planetary-scale waves, are neglected in our model. Note, however, that due to the impact of those stationary forcing mechanisms on the storm tracks, these effects are implicitly incorporated in the cyclone-scale forcing time series.

To test our model hypothesis we are faced first with the problem of identifying blocking episodes in the model flow. In the atmosphere, blocking periods are commonly selected by considering the transient behavior of 500 mbar geopotential height patterns. Therefore, we may identify model blocks from the transient behavior of model flow patterns. For that purpose, we have designed an objective analysis scheme, which virtually combines the techniques of Hartman and Ghan (1980) and Dole and Gordon (1983). The ability of the scheme to identify blocking is tested by applying it to atmospheric data. For the selection of model blocks we have to take into account that the forced planetary modes in the model have lower amplitudes and increased persistence relative to the observed modes (Egger and Schilling, 1984). The blocking activity in the model is examined in terms of statistical distributions of the occurrence of blocks and the structures of mean blocking patterns.

Now, it is evident that the forcing by the cyclone-scale vorticity transfer is only one among a number of possible blocking mechanisms. For an individual case of observed blocking the proposed mechanism may be more or less relevant. It is therefore important to evaluate the contribution of our mechanism to blocking cases that are actually observed in the atmosphere. For that purpose, we compare the mean structure of the model flow for observed blocking with that of the observed flow for the same periods. By that, we hope to find out to what extent the model blocks coincide with the observed blocks.

The concept of stochastically forced flow is reviewed briefly in section 2, where we discuss also the statistical properties of the synoptic-scale forcing. In section 3, we present the blocking analysis scheme and apply it to the observed wintertime flow. We describe the numerical experiments and study statistical characteristics of the forced model flow in section 4. The evaluation of the model blocking activity and its verification against the observed blocking is given in section 5. In section 6 we discuss our results and present our conclusions.

2. Stochastic forcing of planetary flow

Let us first briefly recall the basic concept of stochastic forcing of planetary flow (for a more detailed discussion, see Egger and Schilling, 1983, 1984). The model flow is governed by the equivalent-barotropic vorticity equation on the sphere:

$$\frac{\partial}{\partial t}(\nabla^2 \psi - \Lambda^2 \psi) = -2\Omega a^2 \frac{\partial \psi}{\partial \lambda} + A \nabla^4 \psi, \quad (1)$$

where $\psi$ is the streamfunction, $\nabla^2 \psi$ the vorticity of the barotropic flow, $\Lambda$ a radius of deformation, $a$ and $\Omega$ are the radius and the rate of rotation of the earth, respectively. Here, $\lambda$ is longitude and $A$ is the coefficient of diffusive damping. The effects of orography are not considered in (1). The spectral method (e.g., Elsässer, 1966) is employed for the spatial discretization of (1). The streamfunction $\psi$ is therefore expanded into a truncated series of spherical harmonics:

$$\psi(\lambda, \phi, t) = \sum_{m=-M}^{M} \sum_{n=-n}^{n} \psi_n^m(t) P_n^m(\sin \phi) \exp(i m \lambda), \quad (2)$$

with complex expansion coefficients $\psi_n^m$. The $P_n^m(\sin \phi)$ are the associated Legendre functions of the first kind with degree $n$ and rank $m$; $M$ and $N$ are parameters describing the “basic” truncation of the flow. Furthermore, it is supposed that the model flow is symmetric with respect to the equator. This is achieved by retaining in (2) only those coefficients with odd $n - |m|$. Using the expansion (2), the total flow $\psi$ is partitioned into a planetary-scale regime $\psi_1$ and a synoptic-scale regime $\psi_2$:

$$\psi = \psi_1 + \psi_2. \quad (3)$$

Here, $\psi_1$ (regime I) is defined by the subset of coefficients obtained from a rhomboïdal truncation (R05) of (2):

$$\psi_1 := \{ \psi_n^m; m < 6 \text{ and } n - |m| < 10 \}, \quad (4)$$

while $\psi_2$ comprises all modes in (2) which are not contained in (4). Since the solid-body rotational mode $\psi_0 = -a U_0 \sin \phi$ is conserved by (1) for inviscid flow, it is convenient to write

$$\psi_1 = -a U_0 \sin \phi + \psi_1. \quad (5)$$

Inserting (3) into the vorticity equation yields two separate prognostic equations, one for the planetary-scale flow of regime I and one for the synoptic-scale flow of regime II. For our purposes, only the former is considered. Using (5) and dropping primes, the prognostic equation for the planetary modes reads

$$\frac{\partial}{\partial t} \left( \nabla^2 - \Lambda^2 \right) \psi_1 = \frac{U_0}{a} \left( \frac{2}{a^2 - \nabla^2} \right) \frac{\partial \psi_1}{\partial \lambda} - \frac{2 \Omega}{a^2} \frac{\partial \psi_1}{\partial \lambda}$$

$$- J_1(\psi_1, \nabla^2 \psi_1) + A \nabla^4 \psi_1 - J_1(\psi_1, \nabla^2 \psi_1)$$

$$- J_1(\psi_1, \nabla^2 \psi_2) - J_2(\psi_2, \nabla^2 \psi_1). \quad (6)$$
The first two terms on the rhs of (6) describe wave propagation on the sphere, while the next two terms represent, respectively, the nonlinear interaction between the planetary modes and the effects of diffusive damping. The last three terms represent the nonlinear interaction between the planetary and the synoptic modes. The index "1" on the Jacobians means that only the projection of the interactions onto the planetary modes is retained. Because of the last three terms, Eq. (6) is not closed; they are prescribed as computed from observations. The main hypothesis of ES is to consider this vorticity transfer as a stochastic forcing on the large-scale flow.

In the present paper the cyclone-scale forcing is computed from observations for the 11 winters from 1967/68 to 1977/78. Each winter covers the 96 days from 1 December to 6 March (5 March in leap years). The dataset comprises the Deutscher Wetterdienst Northern Hemisphere 0000 GMT daily analyses of geopotential heights at the tropospheric levels 850, 700, 500, 300 and 200 mb. At each level, streamfunctions are derived from the geopotential heights. A more detailed description of this dataset and the technique used for the solution of the linear balance equation is given in Metz (1985). In particular, we computed streamfunctions in a spherical harmonics expansion with triangular truncation T16 (M = 16, N = 16). Since only Northern Hemisphere data were available, this expansion comprised only the odd (i.e., n = |m| = odd) modes of the spherical harmonics. From the daily streamfunctions the vorticity forcing was evaluated as the sum of the last three Jacobians in (6), using the method of spectral interaction coefficients. Let \( \mathcal{F}(850, t) \) denote this sum of the forcing terms as computed at the 850 mb level at a given time \( t \). The stochastic forcing actually applied is defined as the vertical average

\[
\mathcal{F}(t) = \frac{3}{16} \{ \mathcal{F}(850, t) + \mathcal{F}(700, t) \} + \frac{1}{4} \{ \mathcal{F}(500, t) + \mathcal{F}(300, t) \} + \frac{1}{8} \mathcal{F}(200, t).
\]

Each day the forcing \( \mathcal{F}(t) \) is given by 30 complex expansion coefficients. By using a formula in parallel to (7), we also evaluated the vertically averaged streamfunction \( \mathcal{\psi}(t) \).

Consider now the wintertime Northern Hemisphere local standard deviation (rms) of \( \nabla^{-2} \{ \mathcal{F}(t) \} \) (i.e., the forcing expressed as a streamfunction tendency). For its computation we have first recovered daily maps of this term at a \( (10^\circ \times 5^\circ) \) longitude-latitude grid. At each gridpoint the variance of the daily values was evaluated for each winter separately. The grid values obtained from the individual winters were averaged and the square root was taken. Figure 1 displays the resulting field on a Northern Hemisphere stereographic map. We recognize maxima of wintertime rms over the Atlantic and North America, in accordance with the regions of maxima of low-frequency variance computed by ES for two calendar years. The maximum rms in Fig. 1 is in the Western Atlantic near 60\(^\circ\)N, i.e., just north of the Atlantic maximum of the synoptic regime 500 mb geopotential height rms presented by Blackmon and White (1982). Their Pacific maximum, however, has no counterpart in the forcing rms pattern of our Fig. 1. We come back to this feature in the last section.

Egger and Schilling (1984) have shown that in the atmosphere a prominent part of the planetary modes experience the vorticity transfer from the synoptic scales as a kind of stochastic forcing. They were able to demonstrate that the quasi-stationary and retrograde modes of the flow exhibit statistical characteristics that agree well with the theory of stochastically driven flows. In particular, they considered the covariances between the forcing and the planetary modes. They found both theoretically and from observations that these covariances are small and negative before the forcing has acted. However, near zero lag, the covariances become positive and attain maximum values one day after the forcing has acted.

As the Fourier modes do not give direct information about the regional distribution of this correlation, we have reexamined Egger and Schilling's results in physical space. For that purpose we have prepared maps of the local lagged-correlation coefficient \( r(\lambda, \phi, \kappa) \) between \( \mathcal{\psi}(t) \) and \( \nabla^{-2} \{ \mathcal{F}(t) \} \) for the 11 winter seasons. Figures
2a–c display this correlation at lags of \( k = -1 \) day, \( k = 0 \) days and \( k = 1 \) day, respectively. The local 95\% significance limit for the correlations is 0.1. (Actually, this is an upper limit to a spatially varying significance level; see Appendix). At \( k = -1 \) day (Fig. 2a) we have small negative correlations of order \(-0.1\) over a large part of the hemisphere. At zero lag (Fig. 2b) there are large areas of positive correlations of order 0.1. Considerably stronger positive correlations, however, are found at a lag of one day, i.e., just after the forcing has acted upon the planetary flow. The greatest correlations (0.38) occur over the Atlantic, just where the maximum in the rms pattern of Fig. 1 is observed. Correlations are weaker over the Pacific. At lags \( k > 1 \) day (not shown) the correlations again decrease.

The dependence of the local correlations on the lag corresponds closely to the findings of Egger and Schilling for the slow Fourier modes. From this we conclude that the total planetary flow is indeed influenced by the stochastic forcing from the cyclone scales. It is most interesting that this impact is strongest over that part of the Atlantic Ocean where blocking highs are most frequently observed. Thus, with respect to the main topic of the present paper, the cyclone-scale forcing...
appears to be a promising candidate for the explanation of blocking highs.

3. Identification of blocking episodes

The main object of the present paper is to investigate the blocking activity occurring in a model flow with a synoptic-scale forcing mechanism. To verify the model blocking activity against observations, it is desirable to use the same procedure for the selection of model blocks as of observed blocks. Therefore, we develop in this section an objective scheme for the identification of blocking high episodes.

We consider the barotropic mode of the flow, as given by the vertically averaged streamfunction $\psi$ in a spherical harmonics expansion (2) (cf. section 2). Daily gridpoint values of $\psi$ are recovered in a $10^\circ$ longitude by $5^\circ$ latitude grid for the 11 winters of our dataset. The grid covers the latitude belt from $40^\circ$ to $75^\circ$N, i.e., the latitudes where most of all blocking highs are observed (e.g., Treidl et al., 1981). From the “raw” gridpoint values we evaluated local streamfunction anomalies. For that purpose, we computed at each gridpoint a winter-mean time series by averaging values on corresponding calendar dates for the 11 winters. A long-term trend was defined as the least-squares quadratic fit to this winter mean time series. Local daily anomalies were evaluated as the daily deviations of the raw gridpoint values from the long-term trend.

Cases of blocking are identified by tracking the centers of highs of local streamfunction anomalies similar to the procedure of Hartman and Ghan (1980). (Note, however, that Hartmann and Ghan, in contrast, analyzed raw gridpoint values of geopotential heights.) We started a case if the anomaly in the center exceeds a certain threshold value $\psi_{\text{crit}}$. It is ended if 1) the amplitude falls below the threshold value, or 2) the displacement of the center within one day is greater than $20^\circ$ longitude, or 3) the high leaves the $40^\circ$ to $75^\circ$N latitude belt. We recorded only those cases which lasted for at least five consecutive days. Note that these criteria do not imply that a selected case is associated with a quasi-steady blocking high. To distinguish between blocks and transient ridges Hartman and Ghan selected only the more intense highs as blocks. Because of the differences between our analysis and that of Hartman and Ghan, we reexamined the relation between intensity and duration of the highs.

Using a threshold value of $\psi_{\text{crit}} = 15 \times 10^6 \text{ m}^2 \text{ s}^{-1}$, 143 cases of persistent highs were selected in our analysis. A scatter diagram between the intensity of the highs (mean amplitude averaged over the life cycle of the case) and the duration of the cases is displayed in Fig. 3a. One recognizes only a weak correlation between amplitude and duration. Looking for a more robust relation, we have computed additionally the mean longitudinal displacement rate of the centers of positive anomalies. The scatter diagram between this variable and the duration is shown in Fig. 3b. One finds that

**Fig. 3.** Scatter diagrams for high anomalies selected from observed flow (143 cases, see text for explanation). (a) Mean amplitude vs duration, (b) mean displacement rate vs duration and (c) mean amplitude vs mean displacement rate. Dashed lines in (c) indicate the threshold values applied to associate a case with a case of blocking.
the majority of the highs are prograde and that the longer-lasting cases tend to move slower. Note that the scatter diagram between the mean amplitude and the displacement rate (Fig. 3c) shows only a weak tendency that slower cases also exhibit at larger amplitudes. On the other hand, there also exists a number of intense rapidly moving highs, which obviously cannot be eliminated by an intensity threshold alone. We therefore decided to identify selected cases with blocking events only if the absolute value of the displacement rate of a case was less than 5° longitude per day. Additionally, we also excluded the very weak cases with mean amplitudes of less than $20 \times 10^6 \text{ m}^2 \text{s}^{-1}$. The remaining 88 cases (framed by dashed lines in Fig. 3c) were associated with blocks.

We consider now the regional occurrence of the selected blocks. In Fig. 4 the case averages of the geographical positions of the centers of the blocks are marked. Most cases occur over the Atlantic and Pacific oceans, respectively. A third region of relatively frequent blocking exists in the northern Soviet Union, whereas virtually no blocks occur over the American continent. These features agree well with visually analyzed blocking statistics (Treibl et al., 1981) and with the regions of persistent high anomalies found by Dole and Gordon (1983). To estimate the significance of these results we prepared frequency distributions of the occurrence of blocking by longitude (Fig. 5a), by latitude (Fig. 5b) and by duration (Fig. 5c). The blocking frequency by longitude exhibits the mentioned prominent peaks over the oceans. In terms of a chi-square test this distribution is significantly different from a uniform distribution. The latitudes of most frequent blocking are between 45° and 60°N with a maximum at 55°N. Recall, however, that our analysis procedure cannot detect cases north of 75°N or south of 40°N.

Figure 5c shows that blocks with a duration of 5–6 days are the most frequent events. This finding is in slight disagreement with visually analyzed blocking statistics, which show that the most frequent blocking durations are 8–10 days. This disagreement may be explained by a tendency of our analysis scheme to break longer-lasting blocks into shorter pieces. To test the sensitivity of the analysis scheme to the threshold value we repeated the whole analysis with a larger threshold value of $20 \times 10^6 \text{ m}^2 \text{s}^{-1}$. By this analysis 66 cases of blocking were selected. It turned out that the shape of the resulting frequency distributions (not shown) remained virtually the same as in Fig. 5, indicating the statistical stability of the distributions.

To get an impression of the spatial structure of the selected blocks, we evaluated blocking mean stream-

**Fig. 4.** Mean positions of the observed blocking highs (88 cases).
function anomalies. We averaged the daily streamfunction anomalies over all days on which our analysis identified a blocking case in a prescribed sector. In particular, the blocking mean anomaly streamfunction pattern was evaluated for blocks occurring from 170°E to 160°W over the Pacific and from 40° to 10°W over the Atlantic, respectively. Thirteen blocking episodes occurred over the Pacific and 12 over the Atlantic (cf. Fig. 5a). The mean anomaly pattern for the Pacific blocks is displayed in Fig. 6a. Shaded regions indicate statistically significant mean values (95% t-test). The effective number of degrees of freedom (ε) was estimated as ε = D/D₀ where D is the total number of blocking days in the respective sectors and D₀ (=5 days) represents the estimated time between effectively independent samples (Leith, 1973). The Pacific blocking anomalies are dominated by a pronounced high at 50°N, 180°W. Note that there are also significant lows downstream and to the south of the blocking high. The mean anomalies for Atlantic blocking (Fig. 6b) exhibit virtually the same structure; however, now the mean blocking high is located at 50°N, 30°W and it is more intense than the Pacific high. A weaker, but also significant anomaly is found over Siberia.

It is interesting to compare these results with the blocking analysis of Shukla and Mo (1983). In spite of major differences in the analysis techniques, it is satisfying to see that our Figs. 6a, b agree in almost every detail with their wintertime composite blocking maps (their Figs. 6a, b). In particular, they also got the result that the Atlantic high is stronger than the Pacific high and that both highs have accompanying lows downstream and to the south. Even the Siberian high anomaly during Atlantic blocking can be found in their analysis.

Finally, we discuss the sensitivity of our analysis procedure against a regime I truncation of the analyzed flow. Such a test is important because in the forced model only the planetary regime I part of the flow is predicted. For this test, we repeated the analysis procedure by using the R05 streamfunction ψ₈ for evaluating the gridpoint anomalies. The selection criteria of the analysis scheme were not changed. The relation between the mean amplitude and the mean displacement rate of regime I highs (not shown) is similar to that of Fig. 3c, although the amplitudes were somewhat smaller and the fraction of rapid eastward-moving highs was reduced. The latter finding reflects the fact that rapidly eastward-moving highs are synoptic-scale features which must be properly distinguished from blocking cases. The distribution of the regime I blocks (75 cases, not shown) by longitude and latitude was virtually identical to the distributions of Figs. 5a, b. The distribution by duration, however, exhibited fewer cases of short-lived blocks (5–6 days), while longer-lived blocks (10–14 days) were now more frequent. We conclude that the elimination of small-scale features tends to lead to an increased persistence of the planetary blocking pattern. We also prepared blocking mean
maps for the regime I anomalies and found that the resulting patterns were identical to that of Figs. 6a, b but with somewhat smaller intensities.

4. Local statistical characteristics of forced model flow

In this section we describe the performance of the numerical experiments. Before we discuss the evaluation of the model blocking activity (cf. section 5), we shall study the statistical characteristics of the model flow in space. This will help us to choose the appropriate threshold parameters for the blocking analysis scheme.

a. Experiments

We performed a number of numerical integrations of the regime I vorticity equation (6); the results of two experiments will be discussed. The first (run 1) uses a linearized version of (6), i.e., nonlinear interactions between the planetary flow modes were omitted [third term on the rhs of (6)]. The equation is then identical to that used by ES. For comparison, we also present a nonlinear run (run 2) where this interaction is allowed for. In both runs we used $U_0 = 11.9$ m s$^{-1}$ for the amplitude of the superrotational wind (the value is the observed wintertime mean); $U_0$ is conserved by the model. A deformation radius of $\Lambda = 1000$ km and a coefficient of diffusion $A = 6 \times 10^5$ m$^2$ s$^{-1}$ were used. From $U_0$ and $\Lambda$ the stationary two-dimensional wavenumber is $8 < n_\theta < 9$. Each model run comprised six winters, where the period of integration for each winter covered 106 days starting on 20 November. The same initial conditions were used for each winter. Run 1 was started with a purely superrotational flow, while in run 2 the initial streamfunction was set equal to the zonal average of the observed wintertime mean. In both runs the cyclone-scale forcing $\tilde{F}(t)$ as given by (7) was updated each day from the observations and then kept constant during the next 24 h of integration (a time step of 6 h was used). Via this forcing time series the model flow may be related to the calendar time. The first 10 days of each winter were considered as a relaxation period; the forecasts for the 96 days on and after 1 December were used for the analysis.

b. Results

We consider first the kinetic energy zonal wavenumber spectrum of run 1. Figures 7a, b display the stationary and transient parts of the spectrum, respectively. There we also give the observed spectra for the total flow (dotted) and the regime I flow (dashed). The overall magnitude of the model spectra was found to depend strongly on the value of the coefficient of diffusion. To compensate for the lack of important energy sources for the planetary waves in the model, we used the somewhat small value of $6 \times 10^5$ m$^2$ s$^{-1}$ for this coefficient. The largest response in the transient kinetic energy for run 1 is at wavenumbers 3–5. Naturally, this is a consequence of the fact that energy-cascade transfer toward larger wavenumbers is prohibited in the linear model. Accordingly, this peak virtually vanishes if the nonlinear interactions between the planetary modes are admitted (run 2, not shown). The standing energy spectrum, on the other hand, is largest at small
The spectrum of stationary kinetic energy indicates that the transient synoptic-scale forcing evokes standing eddies in the model flow, too. They may be of possible importance for the excitation of blocks. Accordingly, in Fig. 9a we show the standing eddies $\bar{\phi}^*$ (the asterisk denotes the deviation from the zonal average) of model run 1. One recognizes a high–low dipole pattern over the Atlantic. However, the positions of the high and the low are just interchanged with respect to a blocking dipole. A high is found over the Pacific at 160°E, which may support blocking in this region. If one compares the model standing eddies with those observed (Fig. 9b), one finds that over the Atlantic the model does reasonably well in reproducing the structure of the observed standing eddies. The model standing eddies have an amplitude of about $\frac{1}{3}$–$\frac{1}{2}$ of those observed (note that the contour interval in Fig. 9b is twice the interval in Fig. 9a). A large discrepancy exists, however, over the Pacific, particularly in the region of the Aleutian low, where the model predicted a stationary high. Thus, we may conclude that the time-mean forcing is not favorable for the development of blocking highs.

An important characteristic of atmospheric flows is their persistence. Egger and Schilling have estimated the persistence $P$ of the planetary modes in terms of the autocorrelation function $r(k)$ as $P = -1/\text{In}[r(1)]$. To get an impression of the local variability of the persistence of the model flow, we computed maps of the autocorrelation $r(1)$ at lag $k = 1$ day. For run 1 (Fig. 10a) $r(1)$ is typically between 0.9 and 0.95 corresponding to a persistence $P$ of 10 to 20 days. No pronounced longitudinal variations of preferred high or low persistence can be recognized. However, $r(1)$ increases toward the pole. The observed autocorrelations for regime I (Fig. 10b), in comparison, exhibit values between 0.8 and 0.9 corresponding to a persistence $P$ of 4 to 10 days, and regions of high persistence are found over the Atlantic and Pacific oceans. Note, however, that the total flow (represented by geopotential height), in comparison, has regions of less persistence over the oceans, particularly near the storm paths (Gutzler and Mo, 1983). The persistence pattern for the run 2 flow is similar to that of run 1. In summary, we find that the forced model flow has about double the local persistence of the observed flow, which agrees with the results of ES for the individual Fourier modes.

Finally, we have examined the correlation between the run 1 model flow and the observed regime I flow. Figure 11 exhibits large positive correlations over the Pacific, over the Northern Atlantic and over Europe. The largest correlations (0.46) occur in the same regions as before in Fig. 2c for the correlation between the forcing and the observed flow. However, the correlations are higher throughout. Thus, the forced model reproduces not only main features of the observed variances and standing eddies, but it also has skill in predicting the variability of the planetary flow, even on a day by day basis. The skill is highest in regions where the observed low-frequency variability of the flow

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**Fig. 7.** Zonal wavenumber spectra of (a) stationary and (b) transient kinetic energy for run 1 (full lines). Observed spectra for total flow (dotted) and regime I flow (dashed).
Fig. 8. Standard deviations for (a) model run 1 flow and (b) for observed regime I flow. Contour interval is $10^6$ m$^2$ s$^{-1}$.

is largest and where blocking is most frequent. We are therefore hopeful that the cyclone-scale forcing will prove to be an important mechanism for the explanation of blocking.

5. The blocking activity in the forced model

Let us now turn to the central concern of this paper: whether the synoptic-scale forcing mechanism can cause blocking of the planetary-scale flow. As the model flow variability has an intensity of about two-thirds of that observed, we use a smaller threshold for the blocking analysis ($10^7$ m$^2$ s$^{-1}$). The larger persistence of the model flow also suggests a longer duration criterion. It turned out, however, that the value of this parameter is not essential for the resulting blocking statistics. Therefore, the previous value (5 days) was retained as it was used for the analysis of the observed flows.

Fig. 9. Standing eddies for (a) model run 1 flow and (b) for observed regime I flow. Contour interval is (a) $2.5 \times 10^6$ m$^2$ s$^{-1}$ and (b) $5 \times 10^6$ m$^2$ s$^{-1}$, respectively.
Scatter diagrams of mean amplitudes versus mean displacement rates for the selected model highs are displayed in Figs. 12a (run 1, 144 cases) and 12b (run 2, 146 cases), respectively. As one had to expect, the amplitudes of the model highs are considerably smaller than those observed. The run 1 cases are concentrated at slow displacement rates with a median at retrograde rates, while the cases for the nonlinear run 2 are much more scattered. This indicates that by nonlinear advection a number of the highs have been moved further away from their starting locations. In comparison to the observations, much fewer rapidly eastward-moving highs were found among the model cases. From the cases shown in the scatter diagrams we selected those with a mean amplitude of more than \(15 \times 10^6 \text{ m}^2 \text{s}^{-1}\) and an absolute displacement rate of less than \(5^\circ \text{ day}^{-1}\) as blocking highs (70 cases for run 1 and 86 cases for run 2, respectively).

Figure 13 displays the mean positions of the run 1 model blocks. Compared to the observed blocks (Fig. 4), the model blocks are more uniformly distributed with longitude. Particularly, a number of cases occur over the American continent where no blocks have been observed in the atmosphere. Considering the distribution of model blocks by longitude (Fig. 14a) two maxima may be identified. Most events occur over the Atlantic, while a smaller peak is found over the Pacific. However, a chi-square test reveals that the distribution by longitude is not significantly different from a uniform distribution. The distribution of model blocks by latitude (Fig. 14b) is similar to that observed, while Fig. 14c indicates that the model blocks are longer-lived than the observed blocks. This feature is readily attributed to the increased persistence of the model flow.

We next examine the blocking statistics for the nonlinear run 2. Here, a larger number of blocks (86) was
selected than in run 1, mainly because the run 2 waves have larger amplitudes. The frequency of run 2 blocking by longitude (Fig. 15a) is larger over the Atlantic than that of the run 1 blocks, while the number of Pacific blocks is unchanged. However, the peak is still not significant. The distribution by latitude (Fig. 15b) suggests that by nonlinearity a number of blocks have been advected southward from their origin. The fraction of long-lived (>12 days) blocks (Fig. 15c) is smaller than for run 1 but still considerably larger than for the observed cases.

So far, we found that the forced model produces anomalous quasi-stationary highs, with some regional preference for the Atlantic sector. To provide further evidence of the significance of the model blocking activity, we examined the spatial structure of the model blocks. In parallel to section 3, we prepared mean maps of the run 1 streamfunction anomalies averaged over all days of Pacific and Atlantic model blocking, respectively. To evaluate the effective number of degrees of freedom for the t-test, we took into account that the characteristic persistence of the model flow (10–20 days) is comparable to the average durations of the blocks. Therefore, one degree of freedom was assigned to each blocking episode. The mean structure of the run 1 Pacific (170°E–160°W) blocks (Fig. 16a) is similar to that of that observed (Fig. 6a). There is a pronounced high in the northern Pacific with an amplitude of about two-thirds of that observed and with the accompanying lows downstream and to the south. This coincidence with the observations becomes even better if one inspects the mean pattern for run 1 Atlantic (40°–10°W) blocking (Fig. 16b). The high has about two-thirds of the observed intensity and it is located at the correct latitudes. Moreover, the downstream low and the low to the south are at their right places. The mean blocking anomaly maps for the nonlinear run 2 (not shown) are similar to that of the linear run, the Pacific high having a smaller amplitude, while the Atlantic anomaly is more intense ($13 \times 10^6 \text{ m}^2 \text{s}^{-1}$) than for run 1.

We have found that the model blocks compare well in their statistical characteristics with the observed blocks, which holds particularly for the Atlantic cases. To evaluate the importance of this cyclone-scale forcing mechanism for observed blocks, we examined the transient occurrence of the model blocks in relation to the observed blocks. For that purpose we have computed maps of mean model streamfunction anomalies, but averaging the model flow over all days when blocking was observed in the reality. Note that the model time is related to calendar time via the time-dependent forcing. The effective number of degrees of freedom is determined in that case by the number of observed blocking episodes in the corresponding sectors. The mean model anomalies for observed Pacific blocking (Fig. 17a) exhibit no significant patterns. The mean model flow anomalies for observed Atlantic blocking (Fig. 17b), however, exhibits a significant high ($55^\circ$N, $45^\circ$W) at the correct latitudes but somewhat to the west of the observed high. It is not surprising that the amplitude is smaller than that of the mean model blocks (Fig. 16), but it has still about one-third of the observed amplitude. The model is correct in predicting the downstream low at $50^\circ$N, $0^\circ$E. There are also negative anomalies to the south of the blocking high, but no closed contours can be identified. It is very satisfying that these findings hold also for the nonlinear run 2. The run 2 Atlantic blocking anomaly pattern (Fig. 17c) with the high at $55^\circ$N, $45^\circ$W and the significant low downstream is nearly identical to that of run 1. Thus we may conclude that the mechanism of barotropic transient cyclone-scale vorticity forcing of planetary flow contributes considerably to the occurrence of observed blocks.
6. Discussion and conclusions

We have examined the hypothesis that transient forcing of the planetary modes by vorticity transfer from cyclone-scale eddies can induce and sustain blocking episodes. First, we showed that the local correlation between the forcing and the planetary waves agrees well with the theory of stochastically forced planetary waves (Egger and Schilling, 1984). We found the highest correlations one day after the forcing has acted. The largest correlations (0.38) occur over the Atlantic near a region of main blocking activity. We then examined the response of the model to the forcing. We found that in the linear and nonlinear runs the model flow exhibited maximal standard deviations at high latitudes in the Atlantic just north of the observed maxima. The intensities of these maxima are about two-thirds of those observed. The results were less satisfactory over the Pacific. Furthermore, we found that the standing planetary eddies which are excited by the time–mean part of the forcing tend to oppose the occurrence of blocking in the model.

It is a very striking property of the linear model flow that it comprises virtually only very slow modes. This is expressed in the very high persistence of the model flow. This feature is readily explained by the theory of stochastically driven flows (Egger and Schilling, 1984). The same is true for the nonlinear run, even if now the persistence is somewhat reduced. We found that the model flow exhibits significant local correlations with the observed flow, particularly over the Atlantic (0.46) and to a lesser degree over the Pacific. Thus, we may conclude that a considerable fraction of the observed transient behavior of the planetary modes of the atmosphere is caused by the interaction with the cyclone-scale eddies.

Let us now come to the central topic of the present paper, namely the forced blocking activity in the model flow. We found that the synoptic-scale forcing induces a number of persistent highs in the model flow. The more intense of these highs were selected as model blocks. It turned out that the model blocks were distributed more uniformly with longitude than the observed blocks. However, a closer inspection revealed two peaks of blocking activity located, respectively, over the Atlantic and over the Pacific oceans. The Pacific peak is less pronounced than that over the Atlantic. It virtually vanishes if nonlinear interactions between the planetary modes are allowed. We also examined the spatial time–mean structure of the model blocking
Fig. 14. As Fig. 5 but for run 1 model blocks.

Fig. 15. As Fig. 5 but for run 2 model blocks.
highs. These structures agree well with the observations for both the Pacific and the Atlantic cases. In this respect our model blocks compare reasonably with the observations of Shukla and Mo (1983).

As a last point we have examined the skill of our model in reproducing the transient occurrence of observed blocks. For that purpose we prepared mean model streamfunction maps as above, but now averaging over all days of observed blocking events. It turned out that the Atlantic mean pattern still remained comparable to those observed, although the amplitude of the mean blocking high was reduced to about one-third of the observed. This finding holds for the linear and the nonlinear run. On the other hand, the model exhibits no skill of reproducing observed blocking episodes in the Pacific.

We conclude that the forcing of the planetary flow by cyclone-scale eddy vorticity fluxes is able to induce and maintain blocking highs, particularly in the Atlantic sector. We are aware, however, that our mechanism is only one among a number of possible blocking mechanisms. Nevertheless, our Figs. 17b, c indicate that this mechanism has considerable significance for the Atlantic blocks. For the failure of this mechanism to explain Pacific blocks, we offer two possible explanations. The first is that Pacific blocks may have different physics than the Atlantic blocks. Observational evidence supporting this possibility is provided by Hansen and Chen (1982), who studied two cases of atmospheric blocking. They found that the Atlantic block was forced by the nonlinear interaction of intense cyclone-scale waves with barotropic ultralong waves, while their Pacific blocking resulted from the baroclinic amplification of planetary-scale waves. Naturally, the latter mechanism is not contained in our model. The second explanation may come from data considerations. Particularly for the earlier years of our dataset, the Pacific region was not very well resolved by the Deutscher Wetterdienst analyses. Thus it may be possible that many cyclone-scale features have slipped through the network of the analysis scheme, so that we are seriously underestimating the forcing by the transient synoptic-scale disturbances. Such a shortcoming may be responsible for the striking differences in the standard deviation of the forcing between the Atlantic and the Pacific oceans (cf. Fig. 1). We therefore plan to reexamine our findings by using a different dataset having a better resolution in the Pacific.

On the basis of our results we suggest that, if one wants to model, e.g., the long-term climate of the planetary flow by means of a simple model, one should use a prediction equation that includes the effects of synoptic-scale eddies. As we have seen (Fig. 16a), the model is able to produce a reasonable mean structure of the Pacific blocks, even if the actual forcing appears to be incorrect. This suggests that, as stated by Egger and Schilling (1984), with respect to a simulation of the long-term climate, it is not necessary to represent every detail of the cyclone-scale fluxes. The model will produce the correct climate only if the statistical characteristics of the forcing are correctly prescribed. As shown by Egger and Schilling this can be accomplished
by fitting a Markov process to the forcing and inserting this Markov process instead of the observed forcing into the model.

Acknowledgments. The author is grateful to J. Egger for suggesting the synoptic-scale forcing as a blocking mechanism and to an anonymous reviewer for his many helpful remarks and suggestions. R. Smith is thanked for carefully reading an earlier version of this paper. This research was supported by the Deutsche Forschungsgemeinschaft (DFG) under Contract Eg 40/8.

APPENDIX

Standard Error of Crosscorrelations

On the zero-hypothesis that two discrete stochastic processes $X^n$ and $Y^n$ have no crosscorrelation, the variance of an estimate $r_{xy}(k)$ of the crosscorrelation at lag $k$ from a finite sample of $N$ pairs $(X^n, Y^n)$ is given by
\[
\text{var}\{r_{xy}(k)\} = (N - k)^{-1} \sum_{n=-\infty}^{\infty} \rho_{xx}(v)\rho_{yy}(v), \quad (A1)
\]

(Box and Jenkins, 1976), where \( \rho_{xx}(v) \) and \( \rho_{yy}(v) \) are the true autocorrelations. Assuming a first-order Markov model for both \( X^n \) and \( Y^n \), the autocorrelations may be written as

\[
\rho_{xx}(v) = \exp(-a_x|v|) \quad \text{and} \quad \rho_{yy}(v) = \exp(-a_y|v|), \quad (A2)
\]

where \( a_x = -\ln[\rho_{xx}(1)] \) and \( a_y = -\ln[\rho_{yy}(1)] \). Inserting (A2) into (A1) and replacing the infinite sums by infinite integrals yields

\[
\text{var}\{r_{xy}(k)\} = 2/(N - k)(a_x + a_y). \quad (A3)
\]

We estimated the variance of the crosscorrelations by (A3), where \( \rho_{xx}(1) \) and \( \rho_{yy}(1) \) were replaced by their sample estimates. In all applications of this formula a sample size of \( N = 1096 \) days was used.

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