Baroclinic waves propagating from a high-latitude source

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SUMMARY

A time-dependent primitive-equation model of the southern hemisphere winter troposphere with an idealized dome-shaped continent centred at 80°S has been used to assess the potential of Antarctica to force stationary waves and to determine the location of the mid-latitude baroclinic storm track. Several integrations of the model for two hundred days were made with the zonal-mean fields fixed at the climatological values. In the time mean of each run an anticyclonic wave response formed over the mountain and a planetary wave propagated into the mid latitudes with substantial amplitude. This suggests that much of the observed stationary asymmetry in the southern, mid and high latitudes may be due to Antarctica.

The response to the Antarctic forcing was different in a steady, linear version of the model. However, interaction between the steady waves appears to be of minor importance to the time-mean pattern. Rather, E-vector fields point to a substantial effect of the transients on the pattern. In particular, the low-frequency transients appear to be important in the balance of the time-mean equations near the mountain.

1. INTRODUCTION

As a major orographic feature, Antarctica can be expected to have an important effect on the general circulation of the atmosphere. Several studies (e.g. Mechoso 1981; Mitchell and Hills 1986; Simmonds and Dix 1986 and Mitchell and Senior 1989) with general circulation models have demonstrated the importance of both the Antarctic orography and the sea-ice when modelling the atmospheric circulation. Substantial changes to both the zonal mean and zonally asymmetric components of the time-mean (steady) wind and temperature fields were reported when either sea-ice or orography was removed from the models—at least for the winter case. Simmonds and Lin (1983) found smaller changes for the January case. In interpreting the atmospheric zonal asymmetry or anomaly the distinction can be made between the local response to the Antarctic asymmetry and the propagation of the asymmetry northwards in the form of Rossby waves (e.g. Hoskins and Karoly 1981 (hereafter HK)). The equivalent barotropic nature of the horizontally propagating component of the steady response allows the waves to be usefully modelled with barotropic models. Using the June–August 1979 time and zonal mean 30 kPa vorticity as the basic state in a barotropic vorticity-equation model, James (1988) found that both the full Antarctic orography and an idealized version of it forced a steady response which propagated into the mid latitudes. The inclusion of nonlinear terms did not greatly change the response. James suggested that much of the observed zonal anomaly in the upper troposphere of the mid and high southern latitudes and, in particular, the split in the time-mean mid-latitude jet near New Zealand may be due to Antarctica.

The validity of the use of such barotropic models for the steady response to Antarctic forcing in a three-dimensional atmosphere is, however, unclear. Vertically propagating waves may be important, at least to the local response. Moreover, the continent is sufficiently high that an assumption of linearity about a zonally symmetric flow may be invalid in baroclinic models. An additional problem may be that the zero in zonal wind near the Antarctic surface produces a singularity in the undamped linear baroclinic equations. The nonlinear and transient flux terms are therefore likely to be important in
the time-mean equations of motion for the atmosphere. In addition, observations (Hoskins et al. 1983) show that zonal asymmetries in the transient fluxes in the mid latitudes affect the stationary waves.

The study therefore addresses the following questions:

(i) Can Antarctica force a long, steady planetary wave in the upper troposphere?
(ii) Will the amplitude and phase differ substantially from the barotropic case?
(iii) How does the distribution of transients respond to the presence of the long wave?
(iv) Will the (zonally varying) transient activity modify the long-wave pattern?

We will not attempt to model realistically the southern hemisphere winter. Rather we will use a spectral, hemispheric, primitive-equation model in which the only forced asymmetry is from the idealized Antarctic continent used by James (1988). The resolution of the model, with triangular spectral truncation at wave number 30 and ten levels in the vertical, is typical of climate models. The horizontal resolution was reduced from that used by James for computational economy. Integrations of the model for several hundred days will be analysed for steady and transient components. A version of the model in which the equations were linearized about the climatological zonal-mean state will also be used. The climatology used throughout the paper, and referred to as the ‘observed’, is based on the initialized analyses produced by the European Centre for Medium-range Weather Forecasts (ECMWF) for the months June, July and August of years 1978 to 1984. An updated version of the climatology was described by Hoskins et al. (1989). The steady solutions to the orographic forcing and the calculations of the unstable normal modes of the linear model are available. These will be compared with the statistics from the integrations of the full equations in order to investigate how nonlinear steady and transient wave interactions affect the time-mean planetary waves.

Here we isolate the modelling of the planetary waves from the problem of the maintenance of the zonal-mean circulation by keeping the zonal-mean quantities fixed throughout the integration. Following James (1988) this will be referred to as the semi-linear model. The propagation of Rossby waves through the observed zonal-mean state can therefore be modelled. This technique facilitates the comparison of the various linear and semi-linear solutions since the zonal mean is specified rather than calculated. We assume that the influence of fluctuations in the zonal mean on the observed seasonal mean waves is in general small. The fixing of the zonal mean may also distort the effect transients have on the steady waves, as will be discussed. Further studies with a general circulation model that produces a realistic zonal-mean state would be needed to assess these effects.

In the next section the baroclinic model that is used throughout the paper will be briefly described. In section 3 the linear theory of waves will be applied to the case of Antarctic forcing. The steady solutions of the linearized baroclinic model for the idealized Antarctic forcing will be compared with the barotropic solutions of James (1988) and with WKB approximate solutions for barotropic Rossby waves. The sensitivity of the solutions to the basic state will be discussed. In section 4 the time dependent, semi-linear integrations of the model will be described and the stationary waves will be compared with the observed waves. The effects of transients on the time mean will be analysed in section 5 and compared with those of interactions of the steady waves. The conclusions of the study form section 6.

2. The baroclinic model

The baroclinic model that is used in this paper is based on that of Hoskins and Simmons (1975) (hereafter HS). It is a spectral representation of the primitive equations
for an adiabatic, hydrostatic and perfect-gas atmosphere with a sigma-level \( \sigma = p/p_s \), where \( p_s \) is the surface pressure) vertical coordinate. A jagged triangular truncation of the horizontal spectrum at wave number 30 is used. We specify that the northern hemisphere is the mirror image of the southern hemisphere so that half the spectral coefficients can be ignored. Ten equally spaced layers in the vertical are represented so that the model variables are vorticity, \( \zeta \), divergence, \( D \), and temperature, \( T \), at \( \sigma = 0.05, 0.15, \ldots, 0.95 \) in addition to the surface pressure.

To represent the turbulent cascade of energy and enstrophy from the smallest resolved scales to the subgrid scales, hyperharmonic diffusion terms were added to the tendency equations of vorticity (\( \dot{\zeta} \)), divergence (\( \dot{D} \)) and temperature (\( \dot{T} \)). Large-scale dampings in the form of Rayleigh friction and Newtonian cooling are included. The additional terms to Eqs. (1), (2) and (3) of HS are:

\[
\begin{align*}
\frac{\partial \zeta}{\partial t} & = -K(\sigma)(\zeta - \dot{\zeta}_R) + \nu \nabla^2 \zeta \\
\frac{\partial D}{\partial t} & = -K(\sigma)(D - D_R) + \nu \nabla^2 D \\
\frac{\partial T}{\partial t} & = -K_T(\sigma)(T - T_R) + \nu \nabla^2 T.
\end{align*}
\]

The surface pressure equation (Eq. (4) of HS) completes the set of equations. A dissipation with a 6-hour e-folding time on the smallest length-scale, \( \nu = 0.386 \times 10^{28} \text{m}^{-2} \text{s}^{-1} \), has been used throughout the paper. The restoration fields (\( \dot{\zeta}_R \), \( D_R \) and \( T_R \)) were fixed functions of sigma and horizontal position. \( K^{-1} \) and \( K_T^{-1} \) are the time-scales associated with Rayleigh friction and Newtonian cooling respectively. Various values of \( K \), \( K_T \) and the restoration fields were used, as will be described. The term in HK allowing explicit dependence of \( T_R \) on \( p_s \) was not used here since it would result in a slightly different wave forcing in the time-dependent model than in the linear model. The forcing of asymmetry due to a mountain is through the term involving surface geopotential, \( \Phi_* \).

The orography used here, and in James (1988), is an idealized Antarctic continent specified by the equation for surface geopotential

\[
\Phi_* = gh_A[1 - \tanh((\alpha - \alpha_0)/\alpha_0)]/2
\]

where \( g \) is the acceleration due to gravity and \( \alpha \) is the angular distance, subtended at the centre of the earth, from the point to the centre of the mountain, \( \alpha_0 \) is a measure of the angular width of the sloping side and \( \alpha_0 \) is the angular radius of the mountain at half the approximate peak height \( h_A \). Here the centre was placed at 80°S, 0°E and \( \alpha_b = 5°, \alpha_o = 10° \) and \( h_A = 3000 \text{m} \). The mountain, illustrated in Fig. 1, is comparable in size with east Antarctica. Away from the mountain the surface is at mean sea level.

The zonal mean of the climatological fields was used as the zonal mean in the model. The \( \sigma \)-coordinate, hemispheric spectral form, referred to for convenience as the model basic state, was calculated as follows. Firstly, for each field and pressure surface the value at each grid point on the globe was replaced by the zonal-mean value, except that the meridional wind field was set to zero. The resulting field was then interpolated to the \( \sigma \)-surfaces assuming the idealized surface geopotential field. The value at each northern hemisphere grid point \( (\lambda, \phi) \) was replaced by the value at \( (\lambda, -\phi) \). The spectral form of each field was then calculated so as to give the best least-squares fit on the grid points.

The zonal wind field was found to be sensitive to the weighting used in the least-squares fit: therefore two versions of the basic state were used. State 1, whose zonal-mean zonal wind is shown in Fig. 2, was obtained by weighting each latitude equally. State 2 resulted from weighting by area, with less weight on the polar region. The
Figure 1. Idealized Antarctic topography. Contour interval 500 m, with shading above 2000 m. Zero contour omitted. Latitudes 89°S, 60°S, 40°S and 20°S are shown.

Figure 2. Observed zonal wind used in state 1. Calculated on sigma surfaces assuming the idealized orography and plotted at mean pressure. Easterly regions shaded. Also shown is the zonal-mean surface pressure.
difference between the winds for the two states has a simple structure, with \([u]_2 - [u]_1\), being positive above \(\sigma = 0.6\) and negative below. The difference peaks at 84°S reaching 8.0 m s\(^{-1}\) at \(\sigma = 0.15\), and \(-4.6\) m s\(^{-1}\) at \(\sigma = 0.85\). State 1 matched the ECMWF field more closely near the pole, while state 2 has stronger easterlies near the mountain surface and a greater shear above to give stronger westerlies at 25 kPa. Results for state 1 will be described in more detail, with those for state 2 used to indicate the sensitivity of the results to the basic state.

3. Linear Calculations

(a) Theory

Consider briefly the theory of orographically forced steady waves with regard to the case of Antarctica. A simple, barotropic, beta-plane model for a response linear in the asymmetry in the orography may be written (Held 1983)

\[
[u] \frac{\partial \xi^*}{\partial x} + \beta v = -re^* - [u]_f h_o \frac{\partial h}{\partial x}
\]

(5)

where \([u]\) is a constant basic-state ‘advecting’ zonal wind in which the waves propagate (with \([\ldots]\) denoting the zonal mean and \(*\) denoting the zonal anomaly or asymmetry), \([u]_f\) a constant ‘forcing’ wind which blows over the topography, \(h\) the height of the orography, \(h_o\) the average fluid depth, \(1/r\) the Rayleigh friction time-scale, \(\xi^*\) the steady-eddy relative vorticity, and \(f\) the Coriolis parameter. Other symbols are used conventionally. The model can be justified if the atmosphere can be considered a fluid layer with depth-independent flow, or if the flow is quasi-geostrophic and equivalent barotropic and so mathematically separable in the vertical. As shown by Held, the latter case is relevant to horizontally propagating Rossby waves in the mid latitudes. In that case the forcing wind should be interpreted as the surface-level wind while \([u]\) is that at the equivalent barotropic level. Since the winds near the surface of Antarctica are easterly, while elsewhere the winds are westerly, even the sign of the most appropriate forcing wind is in doubt in this case. James (1988) argued that the flow above the easterlies may be considered a single barotropic layer.

If the advecting and forcing winds in the barotropic model are the same, the eddy vorticity induced over the mountain is anticyclonic (Holton 1979, p. 89). The major effect of the sign of \([u]\) is away from the mountain, where stationary waves will propagate in westerly flow but not in easterly flow. In the global, barotropic vorticity-equation model of James (1988) both the advecting and forcing winds were the 30 kPa zonal wind, which is westerly outside the tropics. The response over the Antarctic mountain in that model was anticyclonic and a planetary wave propagated into the mid latitudes with substantial amplitude.

In quasi-geostrophic theory (HK) a mountain placed in a mid-latitude westerly airstream produces adiabatic cooling on the upslope which will be balanced by the motion of warm air poleward and advection of temperature by the zonal wind. This results in an anticyclonic peak just west of the mountain ridge, and cool temperatures over the mountain. The longer waves propagate upwards with westward phase tilt while shorter components are vertically trapped. For a realistic mid-latitude wind profile the horizontally propagating response is dominated by the external mode of the system, which is a long Rossby wave. The meridional temperature gradient on which the balance depends is comparable with that in the mid latitudes only on the equatorward flank of the Antarctic mountain considered here. There the surface winds are easterly and the same balance would give cyclonic vorticity over the mountain. In any case, if there is an effective wave forcing, a zonal wave-number-1 external mode with equivalent barotropic
structure should dominate to the north and should shift westward in phase as it propagates northward.

The effect of the change in sign of the mean wind above Antarctica on the forced wave is unclear. In the undamped, three-dimensional equations this sign change produces a singularity or critical line which may lead to the steady solution of a damped model being sensitive to the damping. HK show a linear solution for a mountain at 30°N having an easterly wind in the lower levels over its southern slope. The response near the ground was cyclonic over the equatorward slope and anticyclonic in the westerlies over the poleward slope. At upper levels the response was anticyclonic over the entire mountain. Watterson (1984) used a grid-point model with less damping at shorter scales than in the spectral model of HK. A resonant-like response occurred over Antarctica in winter that was trapped south of 60°S and below 40 kPa. Possibly the critical line and change in sign of gradient of potential vorticity resulted in over-reflection (Lindzen and Rosenthal 1981). The response above the easterlies presumably depends on the way the response propagates through the time-mean critical line. This may be quite unrealistic in a steady, linear model.

In summary, the steady, linear baroclinic response to high-latitude forcing can be expected to exhibit an external mode of long wavelength propagating equatorward. The local response may be cyclonic, but both it and hence the amplitude and phase of the remote wave may be sensitive to the model resolution, damping and basic state.

(b) Solutions to the linear baroclinic problem

In this section we consider solutions to the linearization of the baroclinic model described in section 2 about the zonal mean of the basic state. The method used for finding linear solutions in spectral space is based on that described in Hoskins and Karoly (1981). If the equations are linearized about a zonally symmetric state then the equations, for the complex coefficients for wave number $m$ in the state vector, $X_m$, can be written in matrix form

$$\frac{\partial X_m}{\partial t} = AX_m + F.$$  

The matrix $A$ here includes the effects of dampings and dissipations and $F$ includes terms involving $\Phi_0$, $D_R$, $T_R$ and $T_R$.

The steady, linear solution for the perturbation to state 1 by the asymmetric part $(m > 0)$ of the idealized mountain is shown in Fig. 3. Here the Rayleigh damping has a 1-day e-folding time-scale on the bottom level, 5-day time-scale at $\sigma = 0.85$ and zero damping at higher levels. Newtonian cooling is applied to all levels with a 15-day e-folding time-scale. The standard hyperharmonic dissipation is also applied. The zonal anomaly of vorticity at $\sigma = 0.95$ (Fig. 3(a)) is cyclonic over the mountain. At upper levels (Fig. 3(b)) the pattern is of slightly longer length-scale and the anticyclonic peak near the mountain is shifted to the west.

The long length-scale of the response invites a Fourier decomposition of the form:

$$\xi^o = Re \left\{ \sum_{k=1}^{\infty} a_k(\phi) e^{ik(\lambda + \lambda_k(\phi))} \right\}$$

with $\lambda$ longitude and $\phi$ latitude. The amplitude $a_k$ and phase $\lambda_k$ of the wave-number $k = 1$ component are shown in Fig. 4(a). Note that each wave number is mathematically independent in the linear problem so that this is the response to the wave-number-1 component of the orography. The increase in phase with height (indicated by the counterclockwise rotation of the phase arrows in Fig. 4(a)) between latitudes 80°S and
Figure 3. Linear-model vorticity anomaly for basic state 1 at: (a) $\sigma = 0.95$, extrema $+0.5 \Omega$ at 72°S, 50°W and $-0.4 \Omega$ at 82°S, 10°W and (b) $\sigma = 0.25$, extrema $+0.7 \Omega$ at 82°S, 140°W and $-1.1 \Omega$ at 83°S, 0°E. Contour interval 0.1 $\Omega$, anticyclonic (positive) regions shaded. Latitudes 80°S, 60°S, 40°S, 20°S and the equator shown.
Figure 4. Coefficient of wave-1 vorticity for the linear model, (a) for state 1 and (b) for state 2: amplitude (contour interval 0.5 Ω) and phase (arrow direction, with 0 to the right and increasing anticlockwise). Contour numbers indicated. Regions with amplitude greater than 0.1 Ω hatched. Note that the phase arrow points to the longitude of the ridge if the tail is placed at the pole of a polar projection with the zero meridian to the right.
50°S is consistent with the westward shift of the anomaly. The amplitude structure suggests a slight trapping over the pole by the smaller refractive index at upper levels (Randel 1988), although reflection due to poor resolution is possible. The large amplitude of wave-1 vorticity south of 80°S indicates a strong cross-polar wind, which reaches about 30.9 m s\(^{-1}\) at \(\sigma = 0.25\) where it is towards 65°W. The wave-1 phase structure further north appears to be that of an equivalent barotropic external wave propagating northwards: this comparison will be examined below. The amplitude for wave 1 is greatest at the top level but is relatively weak away from the forcing.

The strong wind at the pole, where the zonal mean wind vanishes, casts doubt on the validity of the wave zonal-mean amplitude separation used in justifying a linear solution. However, a cross-polar wind does not necessarily produce significant nonlinear terms in the equations of motion—as, for example, in the case of a Rossby–Haurwitz wave on a solid-body rotation (e.g. Hoskins 1973) in which the nonlinear term is zero and the linear solution also satisfies the nonlinear barotropic vorticity equation.

The sensitivity to the basic state of the wave-1 response is seen in Fig. 4(b) where state 2 was used with the other specifications unchanged. The cross-polar wind was a third of that for state 1 and in a similar direction, while the mid-latitude response was about half as great. Both responses seem trapped south of 60°S, unlike that in the barotropic result of James (1988). Such a decrease in amplitude for the different states did not occur for waves 2 or 3. The resulting shorter length-scale near the ground for the state 2 solution is seen in the comparison of Fig. 3(a) with the corresponding picture for state 2 (not shown). The cyclonic region over the mountain is shifted a little to the west for state 2. Since state 2 has stronger easterlies near the ground the simple balance, mentioned in part (a), giving cyclonic flow in easterlies is only partially valid. The upper-level patterns for the two solutions are qualitatively similar. The phase change in wave 1 from 50°S to 20°S is approximately 200° to the west at upper levels in both solutions. A westward shift also occurs in the barotropic result. However, the phases of the solutions here are different from the phase of the barotropic result—most strikingly over the mountain.

(c) **Comparison with WKB theory**

The wave-1 response remote from the forcing appears to be an equivalent barotropic Rossby wave. The phase progression of the wave in the mid latitudes is similar to that of James's barotropic result. Hoskins and Karoly (1981) apply WKB ray theory to the linear barotropic equation in order to understand the dispersal of energy in wavetrains which propagate from isolated mid-latitude forcings. Since the remote response here consists largely of a single wave number it is appropriate to compare the response directly with the WKB approximate solution to the linear equation for a stationary wave propagating northwards from a point just north of the forcing.

The WKB solution may be written

\[
\psi(x, y) = \text{Re}\left\{a(y) \exp\left(\int_{y_o}^y i\ell(y')dy' + ip_o + ikx\right)\right\}
\]

where \(\psi\) is the stream function and \(k = 1\) is the zonal wave number. The meridional wave number, \(\ell\), is given by Eq. (5.20) of HK, the positive choice being the wave with northwards group velocity. Here \(a\) is the latitudinally varying amplitude and \(\int_{y_o}^y \ell(y')dy' + p_o\) is the phase of the stream function, the initial value at \(y_o\) being \(p_o\). Mercator coordinates \((x, y)\) have been adopted as in HK. In the undamped case the amplitude determined from the conservation of wave action is inversely proportional to
If we include a damping with time-scale \(1/K\) in the wave equation (as in (1)) and include it at second order in the WKB theory, the wave amplitude is diminished at each latitude by the factor \(\exp(-KT)\) where \(T\) is the time required for the wave to reach the latitude travelling at the meridional group velocity.

It remains for us to chose the equivalent barotropic level. This may be done so that the latitudinal wavelength is like that of the model response. In fact the WKB wavelength for wave number 1 in the mid latitudes is not particularly sensitive to the zonal wind speed. The basic-state wind at \(\sigma = 0.45\) produces a wavelength at 50°S of around 53° of latitude, which is similar to that deduced from Fig. 4(a). Thus we choose as the equivalent barotropic level \(\sigma = 0.45\). The stronger winds aloft in the basic state increase the wavelength by several degrees. For winds at the \(\sigma = 0.45\) level the ‘turning’ latitude for the wave is 70°S and the critical latitude is 12°S. The initial latitude here is 68°S which is near the mountain edge.

The amplitude and phase of the complex coefficient for the wave-1 zonal wind from the WKB solution are shown in Fig. 5 for both the undamped wave and for a wave damped with a 5-day time-scale. The zonal wind has been calculated by differentiation of the stream function. The initial amplitude and phase of the stream function is arbitrary but has been chosen to be representative of the northward component of the observed stationary wave, as will be described in section 4. The wave-number-1 component of the zonal wind at \(\sigma = 0.45\) in the linear response for state 1 is shown in Fig. 5 for comparison.

The phase progression in the linear solution is fairly close to that of the WKB solution north of 50°S. This shows that there is little southward wave component in the linear solution and hence little reflection from the critical latitude, consistent with WKB theory (HK). The rapid decline in wave amplitude near 66°S in the WKB results occurs a few degrees to the north in the linear result. In the mid latitudes the amplitude in the linear result appears to be less damped than the damped WKB result. However, in the linear result the fluctuations in amplitude there do not match the small variations in the WKB results owing to the varying meridional wavelength. In fact the WKB amplitude for the zonal wind depends, particularly near the turning latitude, on the way it is calculated from the stream function, in particular whether \(f\) is ignored (as is assumed at first order in the WKB theory) or not. This dependence is more marked for vorticity. Evidently the amplitude of the differentiated fields of the WKB solution are not reliable in this case. One can only conclude that the moderate fluctuation with latitude in the amplitude of the wave-1 component of vorticity seen in Fig. 5 is not necessarily inconsistent with the mid-latitude response being a Rossby wave. The close similarity of the two phase progressions suggests that the linear response is well described as a slightly damped northward-propagating Rossby wave.

4. SEMI-LINEAR CALCULATIONS—THE LONG-WAVE PATTERN

(a) The time-dependent model

James (1988) concluded that in the nonlinear barotropic model, in which the zonal-mean vorticity was relaxed towards the observed 30kPa field, the steady waves forced by the idealized Antarctic orography were weakly affected by wave–wave interaction and that wave–mean flow interaction was unimportant. This small effect of nonlinearity is less likely to be true near the surface of the baroclinic model where the zonal mean winds are weak. Unfortunately the steady, nonlinear baroclinic problem at the resolution of the model used in section 3 has not been solved. Simply running the time-dependent model forward in time does not produce a steady solution (unlike the barotropic case)
Figure 5. Wave-1 component of zonal wind; (a) amplitude and (b) longitude of ridge (− phase for wave 1), for: (i) WKB solution with zero damping (circles) and 5-day damping (dashed), scaling chosen as in text, (ii) linear model with state 1 at $\sigma = 0.45$ (thin line) and (iii) stationary wave in run 1 at $\sigma = 0.45$ (thick line) (see section 3).
because the calculations are dominated by transients in the baroclinically unstable mid latitudes. These transients can also change the time-mean flow. Provided the model can be run for a climatological period (at least several hundred days) the effects of both nonlinearity and transients together can be seen. Such experiments will be described in this and the following section.

Clearly it is desirable that the time- and zonal-mean fields of the model climatology remain close to the observed winter values. However, the model does not include important physical processes, in particular radiative forcing, which produce the observed structure. Radiative forcing of waves, which James (1988) finds to be potentially important, is also ignored. Since we are interested in the planetary waves here we shall fix the zonal-mean fields at their observed values. We therefore ignore the potential effects of zonal-mean variability on the steady waves. This effect may be imposed either directly through wave-transient zonal-mean terms or indirectly through a distortion of the transient eddy statistics.

A possible alternative to the fixing of the zonal means is the use of relaxation terms, similar to the damping terms in Eqs. (1) to (3), which would force the time-mean fields towards realistic values. This has not been attempted owing to the difficulty of determining a form of the terms, if indeed one exists, which would be effective in all the cases considered. In effect the method used here is equivalent to a relaxation of the zonal means towards the observed with a very short time-scale.

It was unclear a priori how the transients resulting from baroclinic instability would behave in the model. In the various experiments or runs of the model that will be discussed here it was found that the magnitude of transient kinetic energy was dependent on the damping used in the model. For the damping used in section 3 the globally averaged energy and also the latitudinal and vertical distribution of the energy were sufficiently realistic that the model may be useful for demonstrating the possible effects of transients on a steady wave. It is, however, possible that this effect is influenced by the fixing of the zonal means.

In the experiments described here the $m = 0$ coefficients of the state vector were returned to their original values at the end of each time-step. The dampings, restoration fields and zonal-mean states of the runs mentioned are shown in Table 1. The experiment labelled run 1, for the more realistic state 1 and moderate damping, will be examined in detail.

<table>
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<th>Run</th>
<th>Description</th>
<th>Basic state</th>
<th>$T_R$</th>
<th>$K$</th>
<th>$K_T$</th>
<th>EKE</th>
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<td>1</td>
<td>Moderate damping</td>
<td>1</td>
<td>1</td>
<td>$8^\approx$, 5.1</td>
<td>10$^3$15</td>
<td>0.94</td>
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<tr>
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<td>Low damping</td>
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<td>2</td>
<td>$8^\approx$, 15.5</td>
<td>3.87$^3$15,2</td>
<td>3.10</td>
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<tr>
<td>3</td>
<td>Moderate damping</td>
<td>2</td>
<td>1</td>
<td>$8^\approx$, 5.1</td>
<td>10$^3$15</td>
<td>0.64</td>
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<td>Continuation of 3</td>
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<td>5</td>
<td>As 3 but half orographic asymmetry</td>
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<td></td>
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<td>0.78</td>
</tr>
</tbody>
</table>

For $T_R$, field 1 is zonally symmetric on each sigma surface, field 2 is as 1 but the bottom-level temperature includes wave components leading to symmetry on pressure surfaces. In all runs the restoration vorticity and divergence fields are symmetric on sigma surfaces. Days 76 to 275 of each run are used in the analysis except for run 4 where days 276 to 475 are used. For $K$, $K_T$, damping times in days from top to bottom levels are listed, with $n^*$ meaning $n$ repeats and $\approx$ no damping. EKE is the time and global average of total eddy kinetic energy, in units of 10$^3$ J m$^{-1}$.

For run 1 the model was integrated for 275 model days starting from an initial state consisting of the zonal mean of state 1 plus a small white-noise wave perturbation in surface pressure. The asymmetry in the orography also produces growth in the waves.
Here the global mean eddy kinetic energy grew steadily to a peak of around $10^6 \text{ J m}^{-2}$ at day 40. From then until day 250 the energy varied between $0.4 \times 10^6 \text{ J m}^{-2}$ and $1.3 \times 10^6 \text{ J m}^{-2}$. Between days 250 and 270 the value peaked briefly near $2.0 \times 10^6 \text{ J m}^{-2}$. Analysis of this peak and similar ones in other runs showed that most of the energy of such peaks was in waves 1 and 2 and in the upper three levels. These peaks are likely to be unrealistic since an unconstrained mean flow would be altered by the large nonlinear terms in a way that would affect the growth of the waves. The peaks dictated that 48 time-steps per day be used to avoid computational instability (the CFL condition). The simple test for static instability of comparing potential temperature between adjacent levels showed that there were frequent occurrences throughout the run of a region slightly warmer than that above, especially at the lower two model levels. Frederiksen (1981) noticed static instability occurring before his model became unstable at day 55, but it did not seem to correspond to high energy here.

The daily states for the final 200 days of the run were analysed for stationary and transient components. The average eddy kinetic energy during this period was $0.94 \times 10^6 \text{ J m}^{-2}$, as shown in Table 1. The stationary wave contributed only 18% of this energy. The zonal and time mean of the high-pass filtered transient eddy kinetic energy is shown in Fig. 6. The simple filter of James and Anderson (1984) and Lau et al. (1981) was used, with an averaging period of 3 days. The high-pass energy is encouragingly like the observed field of the ECMWF climatology (Hoskins et al. 1989) in shape and amplitude. The peak is, however, shifted some $13^\circ$ of latitude south of the observed peak.

An analysis of the model transients in terms of baroclinic instability is beyond the scope of the present paper. It is worth noting that the most unstable linear mode of the model for wave number 6 (calculated as in Valdes and Hoskins 1988) has a zonal-mean

![Figure 6. Run 1 zonal mean of high-pass filtered (period less than 6 days) transient eddy kinetic energy, unit m$^2$s$^{-2}$. Energy is calculated daily on $\sigma$ surfaces and plotted at the mean pressure of the surface. Also shown is the zonal mean of surface pressure.](image)
energy peak at almost the same latitude as the time-mean peak. An integration of the fixed zonal-mean model, with no topographic asymmetry, perturbed by this mode alone resulted in eddy energy greatly exceeding the run 1 average. Apparently the finite amplitudes of all other modes in run 1 allow each unstable mode to equilibrate at a moderate energy. Equilibration at moderate energy also occurred in a run with fixed zonal means but no topographic asymmetry and with all wave modes perturbed. Gall et al. (1978) also produced somewhat realistic eddy energy in a model with fixed zonal means.

(b) Steady waves

The time-mean (denoted by an overbar) vorticity anomaly at $\sigma = 0.95$ and $\sigma = 0.25$ of run 1 is shown in Figs. 7(a) and 7(b) respectively. The pattern at $\sigma = 0.95$ is quite different from the linear result in Fig. 3(a). Here there is anticyclonic vorticity over the mountain and the pattern has a longer length-scale. A smaller anticyclonic peak emerges at 65°S, 90°W and leads into a spiral outwards corresponding to a westward-progressing wave 1. The pattern also propagates upwards with a westward phase shift which reaches 90° by $\sigma = 0.25$ over the pole. The two peaks in Fig. 7(a) merge aloft and there is thus a slight decrease in length-scale as the asymmetry propagates upwards. The peak anomaly at $\sigma = 0.25$ near the mountain is 0.41 $\Omega$ which compares with about 0.12 $\Omega$ in Fig. 6(a) of James (1988). The extremum at 28°S, 130°E here is $-0.22 \Omega$ but is only $-0.13 \Omega$ at $\sigma = 0.45$. James's peak at 22°S, 10°E is about 0.10 $\Omega$. The response in the baroclinic case is thus a little larger, particularly over the mountain. The pattern in the mid latitudes here is about 60° to the west of that of James. This phase shift is similar to that in the baroclinic model between the lower and upper tropospheric levels above the mountain.

The amplitude and phase of wave-1 vorticity are shown in Fig. 8(a). The peak amplitude at $\sigma = 0.25$ is less poleward than in the linear result. The westward phase change with height reduces to zero here by 60°S compared with 50°S in the linear result. North of 50°S wave-1 vorticity is similar to the linear wave except for a constant factor.

An integration with zonal-mean state 2 was made and is denoted run 3. The average eddy kinetic energy was 32% less than in run 1. Unlike the linear result, the steady vorticity anomaly for run 3 was quite similar to that of run 1. The wave-number-1 component of vorticity is shown in Fig. 8(b) for comparison.

The similarity of the model and the idealized forcing mean that favourable comparisons of the model results with observations may be fortuitous. Nevertheless, comparison of the planetary-wave patterns is appropriate. The model mountain is centred at 0°E but the results are invariant with rotation about the pole. Since Antarctica is centred at around 70°E, we will assume a shift of the observed fields 70° to the west in the comparison.

There is some similarity of the large-scale-model vorticity and zonal wind anomalies to the observed anomalies. At the lower levels the main peaks of anticyclonic vorticity occur over the high ground in both the model and observed fields. While the Fourier representation of vorticity is of limited use near the pole, at 72° the observed wave-1 component, shown in Fig. 9, has a similar phase relation to the topography as has the model field. The phase shift to the west with altitude is seen at that latitude in both fields. Likewise, in both the observed and model fields the wave-1 ridge at 60°S is around 160° to the west of the longitude of the mountain peak. The fields also have similar amplitude at that latitude.

The zonal wind anomaly at 45 kPa for both the model and observed are compared in Fig. 10. The strong positive anomaly at 65°S and 180°E is similar to that 70° to the west in the model field. The stronger winds in the observed over and to the west of 50°S,
Figure 7. Vorticity anomaly in run 1 at (a) $\sigma = 0.95$, extrema $+0.6 \Omega$ at $78^\circ$S, 18$^\circ$W and $-0.3 \Omega$ at $80^\circ$S, 140$^\circ$E and (b) $\sigma = 0.25$, extrema $+0.4 \Omega$ at $80^\circ$S, 40$^\circ$W and $-0.3 \Omega$ at $80^\circ$S, 80$^\circ$E. Contour interval 0.1 $\Omega$, anticyclonic (positive) regions shaded.
Figure 8. As Fig. 4 but for stationary wave in (a) run 1 and (b) run 3.

$80^\circ$E partially match those in the model result. However, the smooth positive spiral connecting these two regions in the model is broken in the observed field. The observed negative anomaly peaks at $42^\circ$S, $172^\circ$E near New Zealand. The corresponding model anomaly near $40^\circ$S, $100^\circ$E is less negative. The model zonal wind at $40^\circ$S increases by about $10 \text{m s}^{-1}$ over $80^\circ$ of longitude to the west, about half the observed change. In the observed (James 1988) this increase, together with the positive anomalies to the north
and south, results in a split in the westerly jet. The split also occurs in the model total wind field but is less pronounced (see Fig. 13(a)).

The wave-number-1 component of the zonal wind at $\sigma = 0.45$ in run 1 was shown in Fig. 5. The phase progression is very like that of the WKB solution and, north of 50$^\circ$S, also like that of the linear solution. The amplitude peaks further north than both the WKB and linear results. The amplitude decline for run 1 in the mid and low latitudes is consistent with a damping time-scale of greater than five days and is similar to that in the linear result.

The sharp change with height in the amplitude and phase of vorticity in Fig. 9 in the subtropics suggests that a substantial fraction of the observed mid-latitude asymmetry may be due to tropical forcing not modelled here. Valdes and Hoskins (1989) find this holds in a linear analysis for the summer case. On the hypotheses that first, wave forcing in the mid latitudes is small (supported, for the case of orographic forcing, by James (1988) and Valdes and Hoskins (1989) for summer), and second, the time-mean responses to forcings at different latitudes add linearly, the observed field may be approximated by a sum of northward- and southward-propagating waves. The northward component may then be compared with the model response in the mid latitudes. This component is estimated here by firstly calculating both northwards and southwards WKB waves, assuming 5-day damping, then scaling the waves so that the observed wave-1 meridional wind at $\sigma = 0.45$ is equal to that of the sum of the two waves at 61$^\circ$S and 46$^\circ$S. The sum is crudely like the observed wave within that latitudinal band. For instance the peak zonal wind amplitude is $5.5 \text{ m s}^{-1}$ at 46$^\circ$S in the sum, while the observed wave-1 amplitude reaches $7.3 \text{ m s}^{-1}$ at about that latitude. The scaling factors are quite sensitive to the damping and endpoints so this method can only give an estimate of the northwards component. The WKB solution in Fig. 5 is this northwards component shifted 70$^\circ$ to the west. The amplitude of the wave is a little smaller than that of the model wave. There is similarity in phase between the two to the north of 60$^\circ$S.
Figure 10. Stationary zonal wind anomaly for (a) run 1 at $\sigma = 0.45$, extrema $\pm 20\,\text{m}\,\text{s}^{-1}$ at the pole and (b) observed at $p = 45\,\text{kPa}$. Contour interval 2.5 m s$^{-1}$, positive anomaly shaded, 1000 m contour of Antarctic orography dashed, outer latitude 20°S.
Regardless of the actual forcing of the southern mid latitudes the amplitude of the model response supports the hypothesis that much of the observed wave in the southern mid and high latitudes, including the split in the westerly jet near New Zealand, is forced by Antarctic topography. The question of the linearity of the addition of responses to several forcings is not addressed here. The time mean of the model integration is, however, quite different from the baroclinic linear result. The combination of nonlinear and transient flux terms changes the local wave response to the high-latitude mountain and also appears to facilitate its propagation into the mid latitudes.

5. THE EFFECTS OF STEADY AND TRANSIENT WAVE INTERACTIONS

The zonal asymmetry in the time-mean state of run 1, presented in the previous section, is quite different in structure over the mountain to the wave in the time-independent linear case of section 3. In the mid latitudes the phase progression of the northward-propagating wave is similar in the two results. However, the amplitude variation with latitude is rather different. These differences are due to a combination of the effects of nonlinear interactions between the steady waves and the effects of transients. In an attempt to distinguish between the effects we will firstly compare the propagating waves in various runs and secondly attempt to analyse the direct effects of the asymmetry in the transients on the time-mean zonal momentum equation for run 1.

A gross assessment of the effects of transients can be made by comparing the steady wave in runs with differing levels of transient energy. For economy of presentation, only the important wave-1 component is illustrated here. In Fig. 11 the wave-1 component of zonal wind at $\sigma = 0.45$ for run 2 is shown. This may be compared with that of run 1 shown in Fig. 4. Run 2 has the same basic state as run 1 but has weaker damping and an average eddy kinetic energy over three times that of run 1. The wave amplitude is nearly as large as before in the high latitudes, but has a different structure. Its amplitude diminishes in the mid latitudes where the phase progression is erratic. While the flow near the ground is anticyclonic, as in run 1, the flow across the pole at upper levels is directed 40$^\circ$ to the east of that in run 1. The excessive transients appear to interfere with both the vertical and northward propagation of the anomaly in this 200-day mean.

As seen in Table 1, run 4 (the continuation of run 3) has lower average eddy kinetic energy than run 3. Run 4 had no large peaks in energy such as occurred at around day 250 in run 3. The time-mean responses to the orographic forcing, shown in Fig. 11, are, however, quite similar. The wave-1 zonal wind in run 4 is a little smaller in amplitude over the mountain, and this reduction is also seen further north. Also shown are the amplitudes of the wave-2 and wave-3 components of zonal wind at $\sigma = 0.45$ in run 4. Near the mountain these components are significant, reaching 2 m s$^{-1}$, but further north they are relatively small. In comparison with run 1 it seems that in this case the steady wave is not sensitive to a change in the average transient eddy energy.

As a test of steady-wave interaction, run 5 was made with the wave components of the orography halved. The inclusion of the zonal mean orography leaves a mountain with an oval shape which peaks at 2060 m at 82$^\circ$S, 0$^\circ$E. A purely linear response to the orographic forcing would leave all wave components of the climatology, including statistics of transients, halved in amplitude and with phase unchanged. The actual steady wave-1 component of zonal wind is quite similar to half those of runs 3 and 4, as shown in Fig. 11. The wave-1 response local to the mountain is apparently the most sensitive to the reduction in the nonlinear terms. This might be expected from the relatively large amplitudes of wave 2 and 3 in run 4 over the mountain.
Figure 11. Wave-1 component of zonal wind at 45 kPa; (a) amplitude and (b) longitude of ridge (− phase for wave 1) for: (i) run 2, low-damping case (thick lines), (ii) run 3, full mountain (dashed), (iii) run 4 (thin lines), (iv) run 5, half-mountain case (circles). Amplitude of wave 2 (triangles) and wave 3 (dots) for run 4 also marked.
Chen and Trenberth (1988) found that the change to the wave forcing by topography on the inclusion of terms involving the interaction of eddies with the topography resulted in a substantial change in the response in their model. These interactions are included implicitly (because of the sigma coordinate) in the semi-linear model, but not the linear model, and must contribute to the change on halving the wave components of the orography. Nevertheless, the steady component does not tend noticeably towards the linear-model response, suggesting the importance of other effects.

The steady wave emerging at 60°S in run 5 has phase some 20° to the west of the full-mountain runs. Thus, if the effects of the transients can be ignored, we would deduce that nonlinearity (which will have a greater effect in the full-mountain case) induces a small eastward shift on the wave-1 component in the mid latitudes. This appears to be opposite to the small shift seen in the barotropic model of James (1988). The effect of the wave–wave interaction on the wave-1 pattern appears to be small in both models.

Despite the unchanged specifications of the damping and the zonal mean in the model, the average transient kinetic energy in run 5 is 20% larger than for run 3. In addition, the zonal asymmetry in the various transient statistics is not consistently halved in run 5 as might be expected from the roughly halved stationary component. These differences in the transients will contribute to the differences in the steady components. In particular, given the larger average energy in run 5, the more rapid decline in the run 5 wave amplitude north of 45°S would be consistent with the decline seen in run 2.

The large local response to the forcing in the linear model suggests that the nonlinear terms evaluated from that response would create a large imbalance near the mountain in the linearized equations. However, from the above comparisons there is no clear evidence that either interaction of the time-mean waves or mid-latitude transients of moderate energy can explain the large difference between the linear and time-mean responses in the baroclinic model. Transients near the mountain need to be examined further.

If the transients are to have an effect on the stationary waves in this model they must have asymmetric statistics. While not the most relevant statistic for this effect, the time averaged transient eddy kinetic energy $0.5(u^*''^2 + v^*''^2)$ (where $'$ denotes the deviation from the time mean) is commonly used to indicate the strength of transients (e.g. Hoskins et al. 1989). The energy at 25 kPa for run 1 is shown in Fig. 12. The distribution is quite variable and is highly asymmetric at high latitudes. An averaging period of 5 days (and hence an approximate cut-off of 10 days) has been used to produce the high-pass energy shown in Fig. 12(b). There is a prominent region of high energy extending half-way around the 57°S latitude circle in which values are nearly double those of the other half. This is coincident with, and slightly downstream of, the zonal-wind jet seen in Fig. 10(a).

A notable feature of the low-pass energy (the difference between the total and high-pass fields) is the pronounced maximum slightly westward of the mountain. The peak is seen at levels 10 to 2, with a slight westward shift with altitude. A similar peak occurs in most other runs. The peak corresponds fairly well with the longitude of northward flow from the pole, especially at $\sigma = 0.25$ where the centre of the steady stream-function vortex is located at around 70°S, 90°E. A sequence of daily maps shows that the vortex moves considerably within the eastern sector resulting in alternately large and small northward winds just west of the mountain. This leads to a large variance in $\sigma$ and hence large low-pass eddy kinetic energy and $\bar{v}^2 - \bar{u}^2$. It is tempting to speculate that the variation may be due to barotropic instability of the northward flow. Turning the usual eastward flow theory (Pedlosky 1979, p. 504) by a right-angle we may expect instability of a northward flow where westward vorticity gradient changes sign, as it does here. The
Figure 12. Run 1 transient eddy kinetic energy at 25 kPa. (a) Unfiltered, interval 50.0 m$^2$s$^{-2}$, values above 200 m$^2$s$^{-2}$ shaded and with peak 500 m$^2$s$^{-2}$, (b) high pass (period less than 10 days), interval 12.5 m$^2$s$^{-2}$ with values above 100 m$^2$s$^{-2}$ shaded.
observed low-pass eddy energy at 25 kPa shows no peak near the pole, but neither is there a significant time-mean cross-polar flow at that level (see also Trenberth 1985). It is possible that the large amplitude of the transients near the pole is the result of the fixing of the zonal means in the model. This possibility needs to be addressed with a fully nonlinear model.

An approximation to the time-mean zonal momentum equation for flow on an f-plane can be written in plane coordinates (Eq. A12 of Hoskins et al. 1983, hereafter HJW)

$$\overline{Du} = f\overline{\theta_{am}} + \nabla \cdot E$$

where the transient forcing is given by the divergence of the 'E-vector' defined by

$$E = (v'^2 - u'^2, -u'v', f\theta'/\Theta_p)$$

where $\theta$ is potential temperature and $\Theta$ a standard vertical distribution of $\theta$. The residual $f\overline{\theta_{am}}$ is due to an effective ageostrophic flow. The fixing of the zonal mean in the model must produce a further zonally symmetric term in the residual here. It may also reduce the validity of Eq. (9). Divergence of the $E$-vector indicates that the transients will tend to increase the strength of the zonal-wind westerlies.

The effect of eddies on the barotropic component of the flow can be emphasized by taking a vertical mean of Eq. (10) through the troposphere. The temperature-flux terms in $E$, at the top and bottom are then small. Hoskins et al. (1989) and James and Anderson (1984) (hereafter JA) present means between 100 kPa and 15 kPa of the observed fields. The barotropic nature of the steady waves in run 1 in the mid latitudes invites a similar analysis.

The vertical mean in the troposphere of the high- and low-pass filtered horizontal components of the $E$-vectors for run 1 is shown in Fig. 13. Here the mean was simply the average of values from levels 2 to 10 with half the weighting on level 2 (in order to produce a 15 kPa to 100 kPa mean). The high-pass vectors here and in the observed (JA) point eastward along the region of high transient energy. The vectors diverge in this region, turning strongly equatorward, hence they tend to reinforce the zonal-wind jet. JA found that direct calculation of $\nabla \cdot E_h$ on the full Gaussian grid reveals a noisy divergence field. Here the acceleration and divergence terms in Eq. (9) have been evaluated from values for run 1 on a grid with spacing approximately 8° in latitude (alternate Gaussian latitudes were used) and 30° in longitude in order to give an indication of the importance of the terms at this length-scale. The divergence of the high-pass vectors in Fig. 12(a) reaches $4.1 \times 10^{-5}$ m$^2$s$^{-1}$ at 69°S, 135°E and $2.7 \times 10^{-5}$ m$^2$s$^{-1}$ at 61°S, 75°E. The vertical mean of the acceleration term $\overline{du/dt}$ reaches $6.4 \times 10^{-5}$ m$^2$s$^{-1}$ at 61°S, 45°E. Thus the high-pass transient terms are significant in the time-mean balance. At 195°E, near the end of the region of high transient energy, the high-pass $E$-vector converges, with divergence reaching $-0.9 \times 10^{-5}$ m$^2$s$^{-1}$ at both 61°S and 69°S. Thus the high-pass filtered eddies reinforce the deceleration at the eastern end of the jet. The time-mean streamlines in the mid troposphere are deflected southward in this region; part of the flow crosses the pole and then heads equatorward across the western flank of the mountain.

The low-pass vectors in the observed, point generally westwards in the mid latitudes but are weak and erratic in the 1982 data (HJW). Here the vector components are quite large, in both the vertical mean (Fig. 12(b)) and at the upper levels, particularly just west of the mountain. In the mid latitudes the vertical mean vectors tend to point more westward than the high-pass vectors, especially over the jet. Near the pole the baroclinic nature of the asymmetry makes the vertical mean of Eq. 10 less useful. Moreover, terms
Figure 13. Run 1 vertical mean below 15 kPa stationary zonal wind (interval 5 m s⁻¹). (a) High-pass (period < 6 days) and (b) low-pass $E$-vectors. Note halved unit vector length in (b).
from both the vertical mean and the 25 kPa equations approximated from the grid of values are very noisy, making their significance hard to assess. There are very high positive and negative values of the divergence of the $E$-vector west of the mountain. It is likely that these are balanced in the time-mean equation by terms involving the strong equatorward wind in that area.

The horizontal high-pass $E$-vector components in the mid latitudes are greatest in the upper troposphere. The horizontal divergence of these components increases with altitude and so they will tend to increase the zonal wind vertical shear, as in the observations discussed in HJW. The poleward heat flux due to transients is strongest in the lower troposphere and hence will tend to reduce the wind shear. The temperature-flux vectors (unfiltered) and the time-mean temperature field at 70 kPa for run 1 are shown in Fig. 14. As in the observed (James and Anderson 1984) the temperature is lowest over the mountain, and heat is transported by the transients toward this region of low temperature. The high-pass vectors (not shown) also point toward the cold region but the latitude of the peak flux is further poleward than that of the unfiltered field. The flux is greatest under the jet.

As indicated by both the magnitude of the transient $E$-vector terms and the change in the model response on the inclusion of time-dependence, the effect on the time-mean asymmetry in the model of the transients is clearly significant. The largest effect is south of 60°S. While the high-pass transients are important between about 60°S and 70°S the more energetic low-pass transients are likely to be the dominant influence, particularly near the mountain.

![Figure 14. Run 1 stationary temperature field (interval 2 K, minimum 230 K) and temperature-flux vectors (unfiltered) at 70 kPa.](image_url)
6. SUMMARY AND CONCLUSIONS

The atmospheric response to an idealized continent centred at 80°S in two versions of a baroclinic model of the southern hemisphere winter has been calculated. In a steady, linear version of the model, predominantly cyclonic vorticity occurred over the mountain. A weak large-scale Rossby wave propagated northward to the subtropics. The remote response was much weaker than that for the steady barotropic model of James (1988). Further the zonal asymmetry was sensitive to the basic state.

The time-dependent version of the model was integrated for several hundred days with the zonal-mean fields fixed, and denoted the semi-linear model. Baroclinic instability resulted in substantial transient energy throughout the period that, however, remained at a realistic level because of the damping applied to the waves. Anticyclonic vorticity occurred over the mountain at low levels, the locus of which shifted westward with height. A long wave propagated steadily northwards with an amplitude a little larger than that of the barotropic result. The phase progression of the wave-1 component was like that of the corresponding WKB solution for a wave at an equivalent barotropic level around 45 kPa, suggesting a good agreement between the model and Rossby-wave theory at this zonal wave number. The barotropic model, therefore, gives a response with similar phase if allowance is made for the approximately 60° westward shift with height above the mountain. In contrast to the baroclinic linear model the stationary-wave pattern was fairly insensitive to the damping and zonal-mean state.

The time-mean asymmetry in the baroclinic model was quite different from the linear result with the wave amplitudes being reduced near the pole and increased in mid latitudes. However, there is little evidence that steady-wave interactions caused the difference. The response to a mountain with halved asymmetry had a structure more like that of the full-mountain response than that of the linear-model response.

Analysis of E-vector and temperature-flux vectors due to transients indicates that transients have a significant effect on the stationary pattern. A mid- to high-latitude storm track occurred in the semi-linear model, located over and slightly downstream of the zonal-wind jet. The high-frequency eddies tend to strengthen the model jet at around 60°S but also decelerate it at its downstream end. The low-pass transient energy peaked just west of the mountain. The large amplitude of the low-pass fluxes near the mountain suggests that they caused most of the difference between the linear and time-mean responses by modifying the flow near the mountain. While the time-mean pattern is the more realistic of the two responses it is unclear whether this effect of the low-pass transients is realistic, since they were more energetic near the pole than are the observed.

The significant amplitude of mid-latitude response to the high-latitude forcing in the model suggests that the zonal asymmetry in the orography of Antarctica may cause a large part of the observed long-wave asymmetry in the mid latitudes as well as the high latitudes of the southern hemisphere. Further investigations with a model that does not rely on the fixing of the zonal means are needed before further firm conclusions can be made.

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