On the forcing of planetary-scale Rossby waves by Antarctica

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SUMMARY

The propagation of Rossby waves forced at high latitudes into middle and low latitudes is studied using the barotropic vorticity equation on the sphere. Ray tracing theory indicates that fundamental differences exist between the response of an atmosphere initialized with typical northern hemisphere winter flow and typical southern hemisphere winter flow to such a forcing, with a substantial response more likely for the southern hemisphere case. Integrations with a fully nonlinear barotropic model confirm this suggestion and indicate that away from the local forcing region, linear theory provides a good description of the response. The most notable feature in the steady solutions is a split of the mid-latitude jet close to New Zealand. The location of this split coincides well with a region where blocking is observed to be prevalent in the southern hemisphere.

1. INTRODUCTION

Studies of the circulation of the southern hemisphere (SH) tropospheric circulation (van Loon and Jenne 1972; Trenberth 1980; James and Anderson 1984; see also the data compilations of Oort 1983; Lau 1984) reveal steady planetary wave structures with substantial amplitudes. The SH waves are about half the amplitude and of generally rather longer wavelength than their northern hemisphere (NH) counterparts.

Figure 1(a) shows the steady components of the 25 kPa streamfunction wave. The predominantly wavenumber-1 character of the wave is clear. The high anomalies across the Tasman Sea and the west Atlantic illustrate the well-known blocking region in the vicinity of New Zealand. This blocking region is more clearly shown by Fig. 1(b), which is a plot of the vertically averaged zonal wind in the troposphere. The tropospheric jet splits south of Australia, resulting in low zonal winds over New Zealand. The zonal wind is a maximum over the Atlantic and Indian Oceans, and weaker over South America.

The large-scale steady asymmetries observed in the troposphere are taken to arise ultimately from asymmetries of the lower boundary, that is, from variations of surface temperature or from mountains. Grose and Hoskins (1979) showed that the major features of the NH upper-level winter flow could be reproduced by a linear barotropic model in which the time and zonal mean wintertime zonal winds at 30 kPa were disturbed by the orography alone. A similar calculation for the SH did not give such a realistic simulation of the observed pattern. Grose and Hoskins comment that the linearization, which implies that fluid parcels are constrained to pass over rather than around the orography, is likely to be particularly poor for Antarctica. At such high latitudes, rather little meridional displacement of the parcels could greatly modify the uplift to which they are subject. The present paper re-examines the Grose and Hoskins conclusions and shows that in some respects their SH results were perhaps better than they suggested. However, a fuller picture of the SH planetary waves requires that thermal as well as orographic forcing be taken into account.

Orography in the SH is dominated by the Andes and by Antarctica. While the mid-latitude jet passes over the southern part of the Andes chain, these mountains apparently play a less important role than do the Rockies in the NH flow. Antarctica itself is a smooth dome rising to over 4 km. It is relatively unaffected by the smoothing necessary before it can be represented in a numerical model of the atmosphere. The summit of the
Figure 1. Mean tropospheric flow for the ensemble of SH winters 1979 to 1984, based on ECMWF analyses. (a) Zonal anomaly of the mean streamfunction $\psi^*$ at 25 kPa; contour interval $2.5 \times 10^6$ m$^2$s$^{-1}$ with negative values shaded. (b) Mean zonal wind $u$ averaged from 15 kPa to 100 kPa; contour interval is 5 m s$^{-1}$ with shading indicating values in excess of 15 m s$^{-1}$.
antarctic plateau is displaced some 10° of latitude from the geographical pole, and the high mountain chains of west Antarctica introduce further asymmetry.

As well as being a major orographic feature, Antarctica is also the major heat sink in the SH. Once more, the sink, being centred over the high plateau, is likely to be displaced from the pole. As well as the generally low insolation and high albedo of such a polar ice sheet, the efficient long-wave cooling of the snow surface to space (especially in the long winter night) gives rise to a strong surface inversion and hence to a remarkably strong and persistent pattern of drainage winds (Schwerdtfeger 1984; Parish 1984). In this way, the orography and the radiative properties of the surface and lower troposphere combine to give what may be an important component of the forcing of the large-scale flow.

![Diagram](image)

Figure 2. Schematic illustration of the conceptual model investigated in this paper. Flow is considered to be barotropic between two isentropes, one of which is just above the boundary layer over the summit of the antarctic plateau while the other is near the tropopause. Drainage flow in the boundary layer can result in vortex stretching in the middle troposphere, while upslope flow results in vortex shrinking.

Figure 2 shows the situation envisaged in this paper in a schematic way. The barotropic approach used by Grose and Hoskins (1979) and Hoskins and Karoly (1981) will be adopted. In fact, the flow is baroclinic; the direction and strength of the wind change especially rapidly with height through the inversion layer. It is supposed that the present arguments apply to a layer of the atmosphere between the isentrope which just touches the top of the boundary layer above the summit of the antarctic plateau and an isentrope near the tropopause. In this way, the potential vorticity concept can be used to justify a barotropic representation of the flow above the boundary layer (see for example Hoskins et al. (1985) for extensive discussions of the applications of potential vorticity thinking in such situations).

The starting point is the barotropic vorticity equation on the sphere which may be written:

\[ D\zeta /Dt = -fD - v \cdot \nabla (h/H) + (\zeta_0 - \zeta) / \tau. \]  

(1)

The term \(-fD\) represents vortex stretching in the mid troposphere. This would arise, for example, from a depth-integrated diabatic cooling. In this particular system, it may be thought of as the effect of convergence above the ice sheet which exists to compensate for the boundary layer divergence as air drains off the orography. Vortex stretching as
air is advected across the orography is described by the term $-fv \cdot \nabla (h/H)$. In a strictly barotropic flow, $H$ is the pressure scale height (around 7.3 km); in this paper, $h_0/H$ is treated as an arbitrary parameter, $h_0$ denoting the maximum height of the orography. As well as measuring the orographic height, $h_0/H$ also describes the ratio between the 30 kPa wind and the lighter wind at the level of the orography. The final term on the right-hand side of the equation indicates a relaxation of the vorticity towards $\xi_o$ on a timescale $\tau$. If $\xi_o$ is zonally symmetric, then the term models both surface drag and the forcing of the zonal mean flow by diabatic and baroclinic processes.

In section 2, ray tracing techniques, based on a linearized form of Eq. (1) with arbitrary but localized forcing, will be used to explore the possible response of the far field to high latitude forcing. Section 3 describes linear and nonlinear integrations of Eq. (1) in which the predictions of ray tracing are tested. The model is then used with more realistic orography to explore the steady response of the SH winter flow to forcing. This study and related sensitivity experiments are described in section 4, where it will be shown that the simple model reproduces certain gross features of the SH circulation remarkably well. Conclusions and possible future extensions of this work are discussed in section 5.

2. RAY TRACING CALCULATIONS

In this section, steady solutions to Eq. (1) are considered when the equation is linearized about the zonal mean flow [a]*. This zonal wind is supposed to be a slowly varying function of latitude. If, furthermore, the forcing is considered to be localized, so that the right-hand side of (1) is zero over most of the globe, ray tracing theory as developed for the barotropic atmosphere by Hoskins and Karoly (1981) may be applied. Ray tracing assumes that wave packets of frequency propagate with the local group velocity, while the spatial structure of the packets evolves slowly in response to the changing local potential vorticity gradients encountered. In this way, a two-space-scale or 'WKBJ' approximation, in which the spatial scale of variations of the potential vorticity gradients is presumed large compared with the wavelength of the Rossby waves, is implicit in the analysis. In practice, it is often hard to justify this assumption; nevertheless, it turns out that the theory can give qualitatively useful results even when the wavelength is long.

The reader is referred to Hoskins and Karoly (1981) for more details of the mathematical development. It proves convenient to transform the linearized version of Eq. (1) onto a Mercator projection of the globe, for which

$$x = a\lambda \quad \text{(2a)}$$

and

$$y = a \ln((1 + \sin \phi)/\cos \phi) \quad \text{(2b)}$$

where $\lambda$ is longitude, $\phi$ latitude and $a$ is the radius of the earth. The local group velocity for steady waves (i.e. $\omega = 0$) is then

$$c_g = (\partial \omega/\partial k, \partial \omega/\partial l) = 2\beta_m k(k, l)/(k^2 + l^2)^2 \quad \text{(3a)}$$

* Throughout this paper, the following conventional notation is adopted: for any quantity $Q$, $\langle Q \rangle$ denotes the zonal average and $Q^* = Q - \langle Q \rangle$ the zonal anomaly, while $Q$ denotes the time average and $Q^* = Q - \langle Q \rangle$ is the anomaly at any instant during the averaging period.
where \( k \) is the zonal wavenumber, \( l \) is the meridional wavenumber and \( \beta_M \) is the potential vorticity gradient on the Mercator map:

\[
\beta_M = \frac{2\Omega}{a} \cos^2 \phi - \frac{1}{\sin \phi} \left( \frac{\partial}{\partial y} \cos^2 \phi \right) (\cos^2 \phi [u])
\]  \hfill (3b)

\( \Omega \) being the rotation rate of the earth. Both \( \omega \) and \( k \) are conserved along a ray, while \( l \) evolves. The value of \( l \) is obtained from the diagnostic relation

\[
l = \pm (\beta_M/[u] - k^2)^{1/2} = \pm (K_S^2 - k^2)^{1/2}.
\]  \hfill (4)

Clearly, when \( \beta_M/[u] < k^2 \), propagation is not possible, and disturbances are evanescent in \( y \). \( K_S \), the total steady wavenumber, plays the role of a refractive index in the ray tracing theory. Rays bend towards maxima and away from minima in \( K_S \).

Grose and Hoskins (1979) and Hoskins and Karoly (1981) used climatological zonal winds from Oort and Rasmussen (1971) as the basis for their calculations. Suitable zonal flow profiles for the present study were extracted from cross-sections of zonal wind analysed at the European Centre for Medium Range Weather Forecasts (ECMWF) for the FGGE year. Data for the 30 kPa level averaged for June–August 1979 (SH winter) and for January–February 1979 (NH winter) were taken. It will be noted that the 1979 SH winter was unusual in the degree to which the tropospheric jet was split. However, results based upon an ensemble of six winters do not differ very substantially from those for the FGGE year. The profiles of \([u]\) and \( K_S \) are shown in Fig. 3. The profiles for the NH are similar to those used in Grose and Hoskins (1979) and Hoskins and Karoly (1981). The SH wind profile shows a noticeable double jet. As a result of the minimum of \([u]\) near 45°, \( K_S \) decreases very sharply poleward of 40° though it increases again to values in excess of three at high latitudes. The minimum of \( K_S \) means that the SH mid-latitude flow prevents all but the longest planetary waves from penetrating from the subtropics to any latitude poleward of 45°S. In the NH case, waves can penetrate to systematically higher latitudes.

![Figure 3](image-url)

Figure 3. The flow profiles at 30kPa and corresponding steady wavenumbers used in the ray tracing calculations. (a) Mean zonal wind \([u]\) for the SH, June–August 1979; (b) \([u]\) for the NH, January–February 1979; (c) the corresponding total steady dimensionless wavenumber \( K_S = a(\beta_M/[u])^{1/2} \) for the SH; (d) \( K_S \) for the NH. Shading in (c) and (d) indicates regions where \( K_S \) is imaginary and so waves are evanescent.
The trajectory followed by a wave packet was plotted by solving
\[ \frac{\partial x}{\partial t} = c_g \] (5)
using a simple 2nd-order predictor-corrector method and calculating the meridional wavenumber from Eq. (4). Some rays for the southern hemisphere are shown in Fig. 4. It is assumed that they emanate from a source in the subtropics, but one may, of course, imagine any point along a ray to be the source of the wave activity associated with that wavenumber. The shorter waves (i.e. \( k \) greater than 3) can only reach rather low latitudes before they are refracted back towards the equator. Longer waves can reach higher latitudes before turning back towards the tropics. In particular, wavenumber 1 can propagate across the minimum of \( K_\Delta \) at 40°S and reach latitudes in excess of 75°S. Such a packet can cross the polar cap rather quickly, requiring only 2–3 days to cross the 60°S latitude circle.

![Figure 4. The trajectories of some rays initiated at 20°S, using the FGGE 30kPa winter flow shown in Fig. 3(a). The crosses mark the location of the wave packets at daily intervals.](image)

As the wave packets return to the tropics, they eventually approach the critical line where \([u] = 0\). From Eq. (4), the meridional wavenumber becomes very large and the group velocity of the packet very small as \([u] \rightarrow 0\). In these circumstances, linear theory breaks down; dissipation and nonlinearity play a crucial role in the wave dynamics. It is often assumed that the critical line will absorb any wave activity propagating into it, though in some circumstances, nonlinearity can result in partial reflection of the packet (Nigam and Held 1983; Killworth and McIntyre 1985). When an ageostrophic circulation \([v]\) is included in the basic state, Schneider and Watterson (1987) show that the latitude where \([u] = 0\) is no longer a critical latitude and some wave activity can be advected across the tropical easterlies.

Similar plots for the NH winter are shown in Fig. 5 (see also Hoskins and Karoly 1981). The principal difference from the SH case is that shorter waves can reach high latitudes, but their speed of propagation is much reduced. Although a long-wave packet initiated in high northern latitudes could eventually reach middle and subtropical latitudes, its slow propagation means that it is likely to be greatly attenuated as a result of surface friction. This contrast between the hemispheres will be the more marked since the mean surface friction is appreciably larger in the NH with its greater area of land than in the SH. A similar contrast between the hemispheres was noted by Webster (1982).
These results must be interpreted with caution, since the WKBJ approximation is certainly not formally valid for very long waves, even if the waves can be treated by linear theory at high latitudes. This reservation is less restrictive than it first appears, since the WKJB approach requires that $l^{-1}$ be small compared with the scale over which $\beta - [u]_y$ changes significantly. For the present, ray tracing is regarded simply as a useful conceptual model of the way in which forcing affects remote regions. Then it can be concluded that a mid-latitude response to high latitude forcing is much more likely in the SH than the NH, and that the response will be in the longest planetary waves. In the next section, integrations of the linear and nonlinear barotropic vorticity equation will be used to test this prediction.

![Figure 5. As Fig. 4, but for the NH winter flow shown in Fig. 3(b), and a source of wave activity at 20°N.](image)

3. LINEAR AND NONLINEAR INTEGRATIONS

In order to confirm the predictions of the ray tracing calculations, numerical solutions of the barotropic vorticity equation (1) are presented in this section. The equation may be written in the form

$$\frac{\partial \zeta}{\partial t} + \frac{1}{(1 - \mu^2)} \frac{\partial}{\partial \lambda} \left\{ U\left( \zeta + \frac{fh}{H} \right) \right\} + \frac{\partial}{\partial \mu} \left\{ V\left( \zeta + \frac{fh}{H} \right) \right\} = \frac{(\zeta_0 - \zeta)}{\tau} + A \nabla^6 \zeta \tag{6}$$

where $\zeta$ is the absolute vorticity, $\mu = \sin \phi$ and $U$ and $V$ are zonal and meridional velocities multiplied by $\cos \phi$. Variables have been made non-dimensional, using $\Omega$ and $a$. This equation was solved using the spectral transform method, in which the vorticity is written as a linear combination of spherical harmonics, with triangular truncation to wavenumber 42. The products $U(\zeta + fh/H)$ and $V(\zeta + fh/H)$ appearing in the advection terms were evaluated in physical space on a grid which was equispaced in longitude and spaced for Gaussian quadrature in latitude, the resolution of the grid being chosen to prevent aliasing of quadratic nonlinear terms. This 'spectral transform' technique was devised by Orszag (1970).

The forcing may be introduced into Eq. (6) in two ways. Either the orography $h(\phi, \lambda)$ may be set to non-zero values, or the effects of diabatic heating may be modelled by introducing fixed divergent wind components $U_a$ and $V_a$, corresponding to the
Figure 6. Showing the steady zonal vorticity anomaly when the nonlinear barotropic vorticity equation model was run to a steady state with forcing by a circular mountain displaced 10° of latitude from the pole. (a) Southern hemisphere case, with the flow relaxed to the zonal flow shown in Fig. 3(a). (b) Northern hemisphere case, in which the zonal flow is relaxed towards the profile shown in Fig. 3(b). The contour interval is $1.82 \times 10^{-5} \text{s}^{-1}$ and anticyclonic values are shaded. The dashed circle near the pole indicates the edge of the mountain. For ease of comparison, both integrations have been carried out on the southern hemisphere.
divergence $D$ in Eq. (1). In the runs to be described, either orographic or diabatic forcing was included in isolation. The terms on the right-hand side of Eq. (6) are dissipations due to surface drag and relaxation to an equilibrium zonal flow, and an internal diffusion of relative vorticity $\xi$ included to absorb any energy cascading to the smallest scales resolved by the model. In all the calculations described in this paper, the relaxation time $\tau$ was taken to be 15 days. The $\nabla^6 \xi$ internal diffusion is highly scale selective; $A$ was chosen to dissipate the $n = 42$ wave on a timescale of 6 hours. It follows that $n = 20$ waves are dissipated on a timescale of 20 days and longer waves are essentially unaffected by the internal diffusion.

Each run was started from an initial state of pure zonal flow, with $\xi = 0$, and integrated in time until the fields became steady. These nonlinear runs were reasonably steady after 40 days of integration. The evolution of the flow consisted of three stages. First, vorticity was strongly modified locally by the forcing. This excited a packet of Rossby waves which, in the case of a high latitude source, propagated slowly across the mid-latitudes and into the tropics during the course of several days. In the final phase, the packet approached the critical latitude where $|n| = 0$; the meridional scale of the packet became very small and internal diffusion ensured that the critical line absorbed such wave activity. Thus, in this third stage, a steady state was established in which the generation of planetary wave activity in the forcing region was balanced by dissipation near the critical latitude.

The character of the calculations to be discussed in this paper is exemplified by a nonlinear run with orographic forcing only due to an idealized isolated circular mountain of radius $10^\circ$ displaced $10^\circ$ of latitude from the south pole along the $0^\circ$ meridian, and for which $h_0/H$ was 0.3. The basic zonal flow was the $30$ kPa FGGE flow for the SH winter, shown in Fig. 3. The form of the response is shown in Fig. 6(a), which is a map of the SH $\xi^*$ for this run. It shows surprisingly good agreement with the ray tracing theory. The response is dominated by zonal wavenumber 1, with just a single trough and ridge at nearly all latitudes. Other wave components are present, as evidenced by the fact that the troughs and ridges are not always $180^\circ$ of longitude apart. The intensity of the disturbances becomes rather smaller in low latitudes, and the meridional scale of the wave shrinks (the increase of scale towards the tropics of the polar stereographic projection used in Fig. 6(a) tends to mask this). Negligible wave activity was able to leak across the equatorial easterlies into the NH.

An experiment using the same mountain, but with the FGGE NH winter basic flow (Fig. 4), is summarized in Fig. 6(b). To ease the comparison with Fig. 6(a), the integration was carried out for SH geometry. The drag timescale and other parameters were identical to those employed in the SH case. Again, the response was predominantly in wave-number 1. But it was confined to higher latitudes, consistent with the very slow propagation (and consequent dissipation) of wave packets predicted by the ray tracing calculations. Furthermore, the lower wind speeds at polar latitudes in this zonal wind profile means that the amplitude of the forcing was also smaller than in the SH run. Both these factors combine so that the effect of high latitude forcing on the NH flow is much less significant than in the SH, a result previous implied by Webster (1982).

Thermal forcing gives a generally similar result. Figure 7 illustrates the steady $\xi^*$ field in a run in which the heating was proportional to $-h/H$. It was scaled so that the implied $D^{-1}$ was 15 days and the forcing was not applied to the $m = 0$ coefficients (it was presumed that the zonally symmetric part of the heating was accounted for in defining $\xi_0$). The maximum thermal forcing was now over the summit of the mountain, while the maximum orographic forcing was over the steepest slope of the mountain. Consequently, the response in the thermal case was fairly similar to that in the orographic runs, but was
displaced some 90° of longitude to the east. By combining the thermal and orographic forcing, it would be possible to adjust the phase of the response between the two extremes illustrated by Figs. 6(a) and 7.

The importance of nonlinear effects for the response was examined by repeating the calculation shown in Fig. 6(a), but using a linearized version of the barotropic model. While not especially efficient, the linear version was entirely consistent with the nonlinear model. It was formed as follows: before evaluating the nonlinear products, all the spectral coefficients with $m > 0$ were multiplied by some small number $\varepsilon$. The resulting contribution to the vorticity tendency was then divided by $\varepsilon$ before calculating $\zeta$ at the next timestep. If $\varepsilon$ is sufficiently small, this procedure ensures that all terms involving products of non-zonal quantities are negligible in the tendencies, while products of a zonal flow and a wave are unaffected. These linear runs were slower to reach a steady state than the nonlinear runs, presumably because there was no scattering of energy in spectral space and some waves were dispersed and dissipated only very slowly. The evolution took the form of a decaying oscillation, with period of about 12 days. Accordingly, the $\tilde{\zeta}^*$ field averaged from day 30 to day 42 is shown in Fig. 8. The general pattern of the field is quite similar to that of Fig. 6(a), but the amplitude, particularly at higher latitudes, is rather greater. Although the linear and nonlinear models were run with identical forcing, the strength of the forcing in the final steady state is likely to be rather smaller in the nonlinear case. Nonlinear barotropic flow over high orography tends to develop in such a way that streamlines pass around rather than over the mountain. The ultimate development of such a flow is the ‘Taylor column’ (see, for example, James 1980) in which no flow can penetrate the region over the orography, and so its effect on the flow is strictly bounded. In these terms, the smaller amplitude of the nonlinear response is to be expected.
Away from the forcing region itself, the positions of the extrema were mostly very similar, suggesting that the pattern of the response is much the same in the linear case (when of course its amplitude is arbitrary) as in the nonlinear. The most notable difference was the position of the maximum at 10°E 20°S in the nonlinear run. This was displaced 30° of longitude downstream in the linear run. Although other extrema were in slightly different locations in the two calculations, the displacements were smaller and did not generally exceed 10° of longitude. This major difference is a result of interactions between the eddies, since the zonal wind and absolute vorticity gradients at that location were very similar in the final state to those in the initial zonally symmetric state for the nonlinear run. This conclusion was checked by running a 'semi-linear' version of the model in which full wave–wave interactions were retained (i.e. ε was set to 1) but wave–mean flow interactions were excluded by keeping all spectral coefficients for m = 0 fixed at their initial values. The results of this calculation were very similar to those of the fully nonlinear run, except immediately over the orography. The extrema of ξ* were in virtually the same locations and of very similar amplitude.

4. EXPERIMENTS WITH MORE REALISTIC FORCING

This section describes runs using a more realistic (though still highly idealized) representation of the forcing processes. The earth's orography, expressed as spherical harmonic coefficients with triangular truncation at wavenumber 42, was used to force a zonal flow based on the FGGE data in nonlinear integrations of the kind reported in the previous section.

The smoothed SH orography is shown in Fig. 9. It rises to 4 km over the Andes and to over 3 km over Antarctica. The highest point on the antarctic plateau is at 80°S 100°E. South Africa has substantial areas about 1 km high, while Australasia has few features
of significance. The heights were scaled by the atmospheric scale height and by an additional factor intended to represent the ratio of the low-level wind to the 30 kPa wind; the total scaling is summarized in terms of an 'effective scale height' of 9 km. With such a scaling, much of the vorticity field away from the immediate vicinity of the orographic forcing was reproduced by a linear model, so the amplitude of the orography is somewhat arbitrary. Attention in this section is confined to the pattern, rather than the amplitude, of the response.

Figure 9. The orography of the SH as represented by a series in spherical harmonics with triangular truncation at wavenumber 42. The contour interval is 1080 m.

Figure 10. The steady zonal vorticity anomaly in the southern hemisphere when the nonlinear barotropic vorticity equation model was run to a steady state with orographic forcing only, based on the real earth orography. Contour interval $1.82 \times 10^{-4} \text{s}^{-1}$, anticyclonic regions shaded.
Figure 11. As Fig. 10, but with forcing only from certain elements of the full southern hemisphere orography.
Figure 10 shows the time-averaged \( \overline{\zeta^*} \) fields for integrations which included orographic but not thermal forcing. South of 40°S, the response is mainly in wavenumber 1. The relatively complicated structure of Antarctica means that the local response is similarly complicated. But north of 70°S or so, beyond the northern edge of the ice sheet, the response is similar to that obtained for the smooth, idealized mountain used in the previous section. This is in agreement with the predictions of the ray tracing calculations, which suggest that the only propagating response to high latitude forcing must be at low wavenumber. Thus the smaller-scale features of the Antarctic continent (such as the Antarctic Peninsula) are expected to affect only the local flow, and not to have any important hemispheric impact. Equatorward of 40°S, the \( \overline{\zeta^*} \) field is dominated by smaller-scale disturbances. An obvious wavetrain can be seen emanating from the Andes, with a zonal wavenumber of around 5, though many other sources of steady wave activity introduce other vortices. Drag acts on the wavetrain, which cannot be identified for more than 180° of longitude. In accordance with ray tracing results, this wavetrain is confined to subtropical latitudes; there is no evidence of waves from the Andes or other mid-latitude features affecting the high southern latitudes.

The origin of the various features in Fig. 10 was revealed by taking the orography of each southern hemisphere continent individually and rerunning the model. The results are shown in Fig. 11. All the runs illustrated in Figs. 10 and 11 were fully nonlinear, and so Fig. 10 is not strictly a linear superposition of the frames shown in Fig. 11. But the results of the previous section suggest that a linear model is capable of accounting for a good deal of the variance in the \( \overline{\zeta^*} \) field, and so such a superposition should be approximately possible in practice. A comparison of the figures reveals that this was indeed the case.

The Antarctic case (Fig. 11(a)) produced a large, wavenumber-1 response poleward of 40°S. The \( \overline{\zeta^*} \) field is very similar to that shown in Fig. 10 at high latitudes. Recalling that the highest point on the plateau is at around 100°E, the pattern is very close to that shown for the idealized mountain at 0°E (see Fig. 6(a)). All other continents give rise to a higher wavenumber response which is confined to the regions equatorward of 40°S. The most important subtropical effects are due to the Andes and to southern Africa (Figs. 11(b) and (c)); note that there is a tendency for destructive interference over southern Africa. These two continents both excite a wavetrain which is confined to the mid-latitudes. Australasia, on the other hand, excites a minor train of waves which propagates more or less directly into the tropics (see Fig. 11(d)). As would be expected from the low surface relief of Australia, the amplitude of this train is rather small; it has no discernible impact on the mid-latitude flow.

The results of section 2 suggest that mid-latitude and subtropical orography could excite long wave disturbances which would affect the polar regions; indeed the WKBJ arguments of Hoskins and Karoly (1981) suggest that if the drag is not too large, the amplitude of the wavetrain will increase towards its poleward limit. The calculations illustrated in Fig. 11 indicate that very little such activity is forced by realistic orography. In part, this is a function of the shape of the mountains. For example, the Andes chain is very narrow; despite its height, it simply does not excite much wave activity on the long zonal scale. A second factor is that any long wave activity excited at low latitudes can only propagate polewards fairly slowly, and thus the surface drag is likely to dissipate the wavetrain. Further discussion of these points is included in section 5.

The significance of the perturbations shown in Fig. 11 is best assessed by comparing them with the basic, zonal mean flow. It will be recalled that there is a degree of arbitrariness in the actual amplitudes of the perturbations, depending upon the value chosen for \( h/H \) in Eq. (1). What is important for the present argument is the relative
magnitude of the response to different forcing regions. Figure 12 shows the $\bar{u}$ field for the flow with full orography (Fig. 12(a)) and with Antarctica alone (Fig. 12(b)). The most prominent feature is the split of the tropospheric jet in the south-eastern part of the Indian Ocean, and the associated 'blocking' region near New Zealand. Both cases show a very similar jet split. The most marked difference in the subtropical jet is the jet maximum downstream of the Andes in Fig. 12(a), a feature which is absent in Fig. 12(b). On the basis of such experiments, it is concluded that the location and strength of the jet split are largely due to the morphology of the antarctic continent.

(a) Full earth orography.

(b) Antarctica only.

Figure 12. Showing the steady state zonal wind for the southern hemisphere when the FGGE winter zonal flow is forced by real earth orography. Contour interval 2.5 m s$^{-1}$; shading indicates values in excess of 25 m s$^{-1}$. 
Mo et al. (1987) carried out a case study of blocking in the New Zealand area, and showed that the inclusion of the antarctic orography had little effect on the blocking. This result is not inconsistent with the conclusion drawn in the last paragraph. According to the present results Antarctica is implicated in the tendency to diffuence upstream of New Zealand. Whether or not this diffuence will amplify into a fully developed block on any given occasion is likely to be more dependent on the local synoptic situation than on remote forcing. It is suggested that the forced planetary wave structure is important in fixing the geographical regions where the tendency for blocking is larger.

The inclusion of heating can modify the longitude of the jet split, as for the runs reported in section 3. A calculation was carried out in which the distribution of the high latitude cooling was coincident with the high orography of Antarctica, while the orographic forcing was set to zero. The response to such forcing was similar to that shown in Fig. 11(a), except that the perturbations were displaced downstream by some 60° of longitude. The effect of the perturbations on the zonal wind field was revealed by a split of the jet, which was well to the east of that seen in Fig. 12. The best fit to observations would be a jet split at some intermediate longitude, a result which could be obtained from a calculation including both orographic and thermal forcing set to suitable levels.

5. CONCLUSIONS

This paper has demonstrated that a simple barotropic model reproduces certain important features of the observed SH winter flow, most notable of which is the split of the tropospheric jet to the south of Australia and New Zealand. Such jet structure is strongly linked to the zonal asymmetry of Antarctica. Detailed comparison of the observed and modelled fields shows that smaller-scale detail in the fields is not well modelled, particularly in the tropics and subtropics. The shortcomings of the model will be discussed below; at this point, the positive achievement in suggesting the dominance of Antarctica as an influence upon the steady mid- and high-latitude flow in the southern hemisphere should be emphasized.

A surprising result of these integrations was that nonlinearity proved not to be particularly important, even with such high latitude forcing. It had been suggested by Grose and Hoskins (1979) that nonlinearity would be a major influence for the planetary wave structure at high southern latitudes. Clearly, any linearization about the zonal mean flow must break down sufficiently close to the pole. But the similarity of the linear and nonlinear calculations suggests that linearity rapidly becomes a good approximation away from the pole. The large-scale, propagating parts of the response are excited around the edge of the continent; according to ray tracing theory, propagating steady waves cannot exist poleward of 75°S. The results reported here imply that planetary wave structures north of this latitude are well described by linear theory, with very little interaction between the planetary waves and the zonal mean flow.

The present work entirely neglects the effects of heating anomalies in the tropics and subtropics. These undoubtedly are extremely effective in exciting fairly short wavelength Rossby waves; one would not expect such disturbances to reach latitudes much higher than 40°S on the basis of the arguments presented in section 3. It is suggested that the zonal variations in the tropospheric jet (especially the maximum near Australia) are closely connected to variations in tropical heating. At higher southern latitudes, the sea surface temperatures are not zonally symmetric and so zonal variations of the heating of the atmosphere are to be expected. However, in mid- and high latitudes, the sea surface temperature distribution is partly, if not largely, a response to the atmospheric flow pattern. The anomalous heating may therefore merely lead to a feedback which could
change the amplitude of the atmospheric response without greatly altering its character. It is not clear how the effects of ocean heating in mid-latitudes could be represented in a barotropic model, and it is likely that further progress in this respect will require the use of a multi-level baroclinic model.

The question of baroclinic effects, while it is extremely important, introduces a number of complications. The vertical shear of the zonal winds at high southern latitudes means that the 30 kPa wind, used for orographic forcing in this work, is very different from the surface wind. Indeed, over the antarctic continent, the surface zonal wind is easterly. The wind vector may turn through nearly 180° through the stable boundary layer. In this paper, the flow on some isentrope above the boundary layer is taken to

Figure 13. Longitude–height sections of the zonal anomaly of mean geopotential height $z^*$, contour interval 25 m, negative values shaded, taken from ECMWF analyses. (a) At 60°S, for an ensemble of SH winters (June, July and August) from 1979 to 1984. (b) at 60°N, for an ensemble of NH winters (December, January and February) from 1979 to 1984.
determine the upper tropospheric circulation. But the question of how the deep tropospheric circulation responds to the presence of the ice sheet requires further investigation with a baroclinic model. Once a disturbance has been excited, its propagation in a baroclinic atmosphere is likely to be rather different from that in a barotropic atmosphere. In particular, wave activity might propagate vertically. There is evidence, as illustrated in Fig. 13, that this coupling between the troposphere and stratosphere is somewhat weaker in the SH; the reduced vertical phase tilt of the steady waves in the NH compared with the NH is one indication of this. The amplitude of the steady waves in the vicinity of the polar night jet of the SH is also smaller than in the NH.

A further complication is apparent when it is noted that a baroclinic atmosphere is intrinsically unstable to wave-like disturbances, as well as supporting the propagation of almost neutrally stable longer waves. This complicates the search for nonlinear steady solutions, which can no longer be approximated by direct integration of the forced equations. The interaction between the baroclinically unstable scales and the propagating long waves is likely to be an important component of the SH circulation system. For example, Hoskins et al. (1983) showed that high frequency transient eddies (which are baroclinically unstable, see Hoskins and Sardeshmukh (1987)) interact with the time mean flow so as to help maintain the diffuence near New Zealand. Hence the effect of Antarctica may be both direct, in forcing a mid-latitude long wave pattern with a New Zealand jet split, and indirect, in that this pattern will steer and distort baroclinic disturbances, thereby amplifying the initial pattern. Shutts (1983) has discussed such a mechanism in an idealized model.

The Reading University multi-level spectral model has been modified to explore the response of a baroclinic atmosphere to orographic forcing at high latitudes. Dr Watterson and I plan to report on these investigations in a further paper which is nearing completion.

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