Orographic versus Thermal Forcing of Stationary Waves: The Importance of the Mean Low-Level Wind

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ABSTRACT

The amplitude of the linear, stationary response to low-level extratropical heating decreases as the magnitude of the low-level mean flow increases, while the amplitude of the orographically forced waves increases. As a result, linear theory predicts that the relative importance of thermal and orographic forcing for the extratropical stationary wave field is very sensitive to the magnitude of the zonal mean low-level winds. In the process of illustrating this sensitivity, we also show how the dependence of the orographic response on the low level winds can be distorted by a numerical σ-coordinate model.

1. Introduction

Linear modeling studies of tropospheric stationary eddies have become more realistic in recent years, yet a consensus on the relative importance of thermal and orographic forcing for the climatological eddies has been slow to emerge. For example, Nigam et al. (1988), attempting a linear decomposition of the stationary eddies predicted by a general circulation model (GCM), find that orography contributes nearly twice as much as does forcing by extratropical heating and transients to the upper tropospheric eddy geopotential in northern winter. In the lower troposphere, the amplitudes are comparable. Yet Valdes and Hoskins (1989), in their study of the observed stationary waves, find extratropical heating to be dominant in both upper and lower troposphere. Part of the discrepancy between the two models is due to differences in the heating fields, but this is not the only cause, since the response to orography obtained by Valdes and Hoskins is smaller than that in Nigam et al. Differences in the prescribed zonal mean basic state must also be important.

Our purpose here is to emphasize that orographic eddies and eddies forced by shallow extratropical heating are both sensitive to the low-level zonal mean winds: orographically forced waves increase and thermally forced waves decrease in magnitude as the low-level winds increase in strength. As a result, one can move from a state on which orographic forcing is dominant to one on which thermal forcing is dominant by reducing the low-level midlatitude westerlies by a modest amount. Orographic and thermal forcing are discussed in sections 2 and 3 respectively. Reasons why one might question the relevance of these linear results for the atmosphere are discussed in the conclusions.

A problem that can arise is using a σ-coordinate linear model to study orographically forced waves is described in the Appendix. To the extent that linear theory is relevant, the implication is that the potential for distortion of the orographic response exists in σ-coordinate GCMs similar to those presently in use.

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2. Orographic forcing

The lower boundary condition for a quasi-geostrophic linear model forced by orography reads, using log pressure as vertical coordinate,

$$f(\bar{u}\partial_z v - v\partial_z\bar{u}) = -N^2 w = -N^2\bar{u}\partial_x h, \qquad (1)$$

where h is the orography, $v = \partial_x\psi$, and ψ is the geostrophic streamfunction. To determine which of the two terms on the LHS of (1) is dominant, one must compare the vertical scale of the forced wave at the surface, $v/(\partial_z v)$, with that of the mean flow $h_u \equiv \bar{u}/(\partial_z\bar{u})$. If an f-plane were adequate for the local response, then $v/(\partial_z v) \approx fL/N$ where L is the scale of the orography. More appropriately for the large scales that dominate the stationary eddy field, we use $\partial_{zz}v = -m^2 v$, where

$$m^2 \equiv N^2(K_s^2 - K^2)/f^2, \qquad (2)$$

$K_s^2 \equiv \beta/\bar{u}$ and K is the total horizontal wavenumber. Effects of compressibility and horizontal curvature of the winds are ignored. We can generally assume that $K_s^2 \gg K^2$ near the surface on the large scales of interest, due to the small mean winds. If the solution close to the source is dominated by upward propagating waves, then we can set $v_z \approx mv$. If downward propagating waves are also significant near the source, this scaling is not legitimate; in particular, the sensitivity to changes in the mean flow could then be determined by proximity to a resonance. We presume that meridional and zonal propagation away from a localized source, combined with dissipation due to drag in the boundary layer and "critical layer" absorption in the tropics, prevents the generation of a significant downward-propagating component in the source region. (From inspection, this is clearly the case in the linear modeling studies mentioned in the Introduction.) In this case, $mh_u \ll 1$ is the condition that ensures that the second term dominates over the first on the LHS of (1), so that the low-level response is directly proportional to $\bar{u}(0)$.

To the extent that WKB is valid, the amplitude of the upward-propagating wave is proportional to $m^{-1/2}(z) \propto \bar{u}^{1/4}(z)$. If one perturbs the mean flow by decreasing $\bar{u}$ by the same amount at all heights, the ratio of upper to lower level amplitudes will increase slightly, and upper level amplitudes will be slightly less sensitive to the mean flow reduction.

In the limit $K \rightarrow 0$, we can write $m^2 = (h_\beta h_u)^{-1}$, where $h_\beta = f^2\partial_z\bar{u}/(\beta N^2) \approx 10$ km in midlatitudes. (If h_β is much larger than the scale height H, it should be replaced by H.) The condition $mh_u < 1$ is then equivalent to $(h_u/h_\beta)^{1/2} < 1$. If we choose $\bar{u}(0) = 5$ m s^{-1} and $\partial_z\bar{u} = 2 \times 10^{-3}$ s^{-1}, then $h_u = 2.5$ km, so this condition is at least marginally satisfied. As mh_u approaches unity, the two terms on the LHS of (1) become comparable, and the orographic response's sensitivity to the low-level winds decreases.

Another consequence of the dominance of the second term in (1) is that the response is inversely proportional to the meridional temperature gradient. The importance of this dependence for an understanding of the stationary waves of the ice age climate is discussed in Cook and Held (1988).

The preceding argument is relevant for the very long waves propagating into the stratosphere, as well as the region near the source where upward propagation dominates. A complementary argument focuses instead on the region removed from the source where reflections within the tropospheric waveguide have set up an external Rossby wave field. The problem in this region is to compute the amplitude of the external Rossby waves radiating horizontally away from the source. In the notation of Held et al. (1985), this amplitude is proportional to $\alpha_s\bar{u}_s$, where $\bar{u}_s$ is the surface wind and α_s is determined by the structure of the external mode eigenfunction. If one fixes the vertical shear and varies only the barotropic component of the flow by varying $\bar{u}_s$, one can think of α_s as a function of $\bar{u}_s$. A plot of this function can be found in Held et al. for several shear profiles (Figs. 7c and 9b of their paper). One finds that α_s approaches a nonzero constant as $\bar{u}_s$ approaches zero, and decreases monotonically as $\bar{u}_s$ increases. As a result, the amplitude of the external mode is proportional to $\bar{u}_s$ for small $\bar{u}_s$, but increases less rapidly as the surface winds continue to increase, a dependence similar to that of the local or upward propagating response.

To check that this dependence on the surface wind is observed in a realistic setting, we use a σ-coordinate primitive equation model on the sphere, linearized about a zonally symmetric basic state. The model is decomposed into spherical harmonics in the horizontal, with wavenumber 15 rhomboidal truncation, and has 10 equally spaced σ-levels in the vertical ($\sigma = p/p_s$ where p_s is the surface pressure). The basic state is taken from an idealized perpetual January integration of a general circulation model with the same horizontal resolution and an all-ocean surface with prescribed zonally symmetric surface temperatures. The orography is a Gaussian mountain centered at 40°N with a half-width of 15° in both latitude and longitude. The dissipative terms included in the linear model are a biharmonic diffusion with coefficient 10^{17} m^4 s^{-1}, a linear Rayleigh surface friction with strength (1 day^{-1}) at the surface, decreasing linearly to zero at 800 mb, and extra dissipation near critical levels as described in Nigam et al. (1986).

The stationary response is computed for a series of basic states obtained by adding the flow $(\lambda - 1)\bar{u}_s(\theta)$ to the mean flow in each layer, where $\bar{u}_s(\theta)$ is the latitude-dependent mean zonal wind in the lowest model layer, and $0 < \lambda < 1.5$. The surface wind is thus modified to equal a fraction λ of its original value, while the vertical shear and temperature field are left unchanged. The dependence on λ of the rms eddy geopotential height at 300 mb, averaged over the region 20°–70°N, is plotted in Fig. 1.

For very small λ the response deviates from that anticipated, in that it does not vanish as the surface wind tends to zero. This result, along with a much larger distortion found when the linear model uses the unequally spaced vertical levels of the GCM, is discussed in the Appendix. Ignoring this distortion, the linearity for small λ and the departure from linearity as λ increases are as anticipated. The GCM's basic state ($\lambda = 1$) has surface winds large enough that the first term in (1) has become comparable to the second, so that the departure from linearity by this point is noticeable. The horizontal structure of the response of this linear model at 300 mb (with $\lambda = 1$) is shown in Fig. 2a. To first approximation, this structure is unchanged as λ and the amplitude of the response vary.

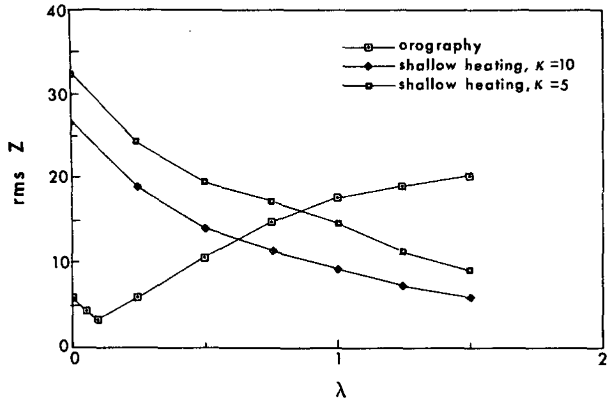

FIG. 1. Sensitivity of the amplitude of orographically and thermally forced stationary waves to the strength of the zonal mean low-level wind, as predicted by a linear model. The rms eddy geopotential at 300 mb is plotted against the parameter λ defined in the text. The maximum amplitude of the mountain is 1 km, while the maximum strength of the heating, averaged in the vertical, is 2.5 K day^{-1}.

A plot of the amplitude at 990 mb as a function of λ is very similar to the results for 300 mb displayed in Fig. 1.

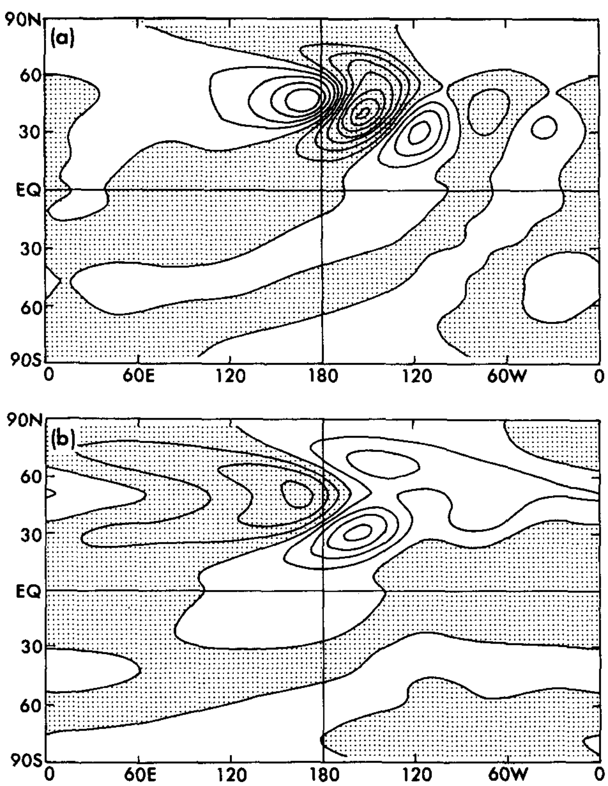

FIG. 2. Horizontal structure of the 300 mb eddy geopotential height forced by (a) orography and (b) shallow heating in the linear model. Shaded areas are negative. For the forcing amplitudes noted in the caption to Fig. 1, the contour interval is 10 gpm.

Figure 3 compares the observed 950 mb winds in the Northern Hemisphere from Oort and Rasmusson (1971) with that produced by the idealized GCM and used as the λ = 1 basic state in this linear model. The GCM's flow is more comparable to that observed in the Southern Hemisphere with its stronger low-level westerlies. The reduction in strength needed to bring this basic state closer to Northern Hemisphere observations would cause a substantial ($\approx$40%) reduction in the response to the idealized mountain in Fig. 2. The GCM analyzed by Nigam et al. (1986) has Northern Hemisphere surface westerlies that are stronger than those observed, lying roughly halfway between the two curves in Fig. 3.

3. Thermal forcing

The response to low-level extratropical heating is conveniently thought of as a sum of particular and homogeneous solutions. The linear, quasi-geostrophic thermodynamic equation reads:

$$f(\bar{u}\partial_z v - v\partial_z \bar{u}) + N^2 w = \kappa q/H \equiv Q, \qquad (3)$$

where q is the heating rate per unit mass, $\kappa = R/c_p$, and H is the scale height. Suppose that the adiabatic cooling due to upward motion is negligible. If the variation of $\bar{u}$ over the heated region is also negligible, i.e., if the vertical scale of the heating h_Q is much smaller than h_u, then we have the simple solution

$$fv_p(z) = \bar{u}^{-1}(z) \int_z^{\infty} Q(\zeta)d\zeta, \qquad (4)$$

where the subscript p refers to the particular solution. If h_Q is not small compared with h_u (continuing to ignore the adiabatic cooling term) the particular solution is only slightly more complicated:

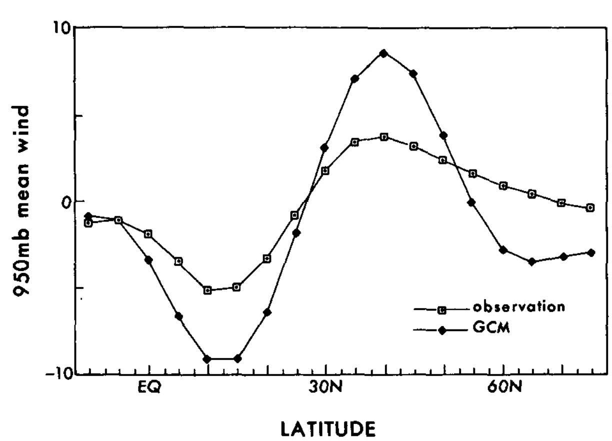

FIG. 3. Latitudinal distribution of the 950 mb mean zonal wind in the Northern winter from Oort and Rasmusson (1971) compared with that produced by the idealized GCM that provides the basic state for the linear model in this study.

$$fv_p = \bar{u}(z)\int_z^\infty Q(\zeta)/\bar{u}^2(\zeta)d\zeta. \qquad (5)$$

For the case of a linear shear flow $\bar{u} = \bar{u}_0 + \Lambda z$ and a heating profile that is constant ($=Q_0$) below the height h_Q and zero above this height, (5) reduces to

$$fv_p = (h_Q - z)Q_0/\bar{u}(h), \quad z < h_Q,$$
$$= 0, \quad z > h_Q. \qquad (6)$$

The response is inversely proportional to the mean wind *at the top of the heated region.* The sensitivity to a depth-independent mean flow modification (of the sort used in section 2) decreases as the depth of the heating increases, since the resulting fractional change in $\bar{u}h_Q$ decreases.

As described in Hoskins and Karoly (1981) or Held (1983), the vorticity equation can be used to compute the vertical motion associated with this particular solution and determine whether or not adiabatic cooling is indeed negligible in (3). For the purpose of these scaling arguments, we consider the separable problem in which $\bar{u}$ is a function of z only, so that

$$\bar{u}(K_s^2 - K^2)v_p = f\partial_z w_p. \qquad (7)$$

Therefore,

$$N^2 w_p/f \propto m^2 h_Q \bar{u} v_p, \qquad (8)$$

where h_Q is once again the vertical scale of the heating. Since $v_p/(\partial_z v_p) \approx h_Q$ for the particular solution, the ratio of the third to the first term on the LHS of (3) is $m^2 h_Q^2$. It follows that the consistency condition is $mh_Q \ll 1$. Since m^{-1} varies from $\approx$3–6 km as $\bar{u}$ varies from 1 to 5 m s^{-1} in midlatitudes, the assumption of small mh_Q is generally useful for qualitative analysis of extratropical heating (see also Webster 1980). If $mh_Q \ll 1$ and $K_s^2 \gg K^2$, then $w_p \propto v_p \propto \bar{u}^{-1}$. For the simple top-hat heating profile, $\bar{u}$ in this scaling relation should be thought of as the flow near the top of the heated region.

To this particular solution one must add a homogeneous solution to satisfy the lower boundary condition $w = 0$. One can think of the homogeneous solution as forced by the vertical velocity $-w_p(0)$. The magnitude of the homogeneous solution near the source, assuming that the second term on the LHS of (1) dominates, is

$$v_h \approx N^2 w_p/(f\partial_z\bar{u}) \approx m^2 h_Q h_u v_p \approx (h_Q/h_\beta)v_p, \quad (9)$$

where the last expression depends on $K_s^2 \gg K^2$. Thus, the homogeneous solution tends to be smaller than the particular solution within the heated region. Above the heated region, v_h is the entire solution. Assuming that $w_p \propto \bar{u}^{-1}$, the arguments of section 2 show that the homogeneous solution is also approximately proportional to $\bar{u}^{-1}$ if $mh_u < 1$.

The amplitude of the external Rossby waves generated by localized heating can be determined by projection, without assuming that $mh_Q \ll 1$. From Held et al. (1985) this amplitude is proportional to $\langle Q\bar{u}^{-2}w_e\rangle$ where w_e is the vertical profile of the vertical velocity in the external mode and the brackets denote a vertical integral. From the vorticity equation we know that $w_e \propto v_e z$ near the surface. But $v_e(0)$ is proportional to $\bar{u}(0)$ when surface winds are small. We thereby regain the result that the response to low-level heating is inversely proportional to the low-level winds.

This analysis of the sensitivity of the extratropical response to prescribed shallow heating is potentially misleading due to the important role played by transient eddy heat fluxes at low levels. The transients act to damp the low-level stationary eddy temperatures, as is clear, for example, from Lau (1979). If we simply add a damping term to crudely represent the effect of the transients, then

$$f(\bar{u}\partial_z v - v\partial_z\bar{u}) + N^2 w = Q - f\gamma\psi_z. \qquad (10)$$

The damping can have a large effect on the particular solution, which has the same small vertical scale as the heating. The homogeneous part of the solution is then affected through $w_p(0)$. The additional direct effect of the damping on the homogeneous solution is relatively small due to its larger vertical scales. (By the same token, the effect of this damping on the response to orography should also be relatively modest.) If the damping effect of transients is a large term in (10), the sensitivity of the solution to low-level winds will clearly be overestimated if one ignores the damping or, equivalently, considers the transient eddy flux convergence as part of the prescribed heating field.

We again use a linear, primitive equation model on the sphere to illustrate the sensitivity of the thermally forced waves to the low-level flow. In addition to the dissipative terms listed in section 2, for $\sigma > 0.8$ we include thermal damping with coefficient $\gamma = (3 \text{ days})^{-1}$. Without thermal damping or diffusion of some sort, we do not obtain a physically reasonable response to low-level extratropical heating in this model. (If we include this thermal damping in the computation of the orographic response, the rms amplitudes in Fig. 1 are reduced by $\approx$10%; otherwise the results are essentially unchanged.) The heating decays away from the surface ($\sigma = 1$) with the form $Q = Q_0 \exp[-\kappa(1-\sigma)]$. Two values of κ are chosen: 5 and 10. The value of Q_0 depends on κ in such a way that the vertically integrated heating is fixed. [The heating in each layer of the finite-differenced model is determined by the integral of $Q(\sigma)$ over that layer.] The location and shape of the heating is identical to that of the orographic forcing. The resulting rms amplitudes at 300 mb are shown as a function of the strength λ of the surface winds in Fig. 1. One sees the expected increase in amplitude as the low-level mean winds decrease. The fractional changes in amplitude are larger for the shallower heating, also as anticipated. One sees a qualitatively similar, but slightly weaker, sensitivity at low levels (not shown). The horizontal structure of the response in 300 mb geopotential is shown in Fig. 2b for the case with $\kappa = 10$. Once again, this structure is insensitive to the value of λ.

As λ increases from 0.5 to 0.75, corresponding to an increase in the strength of the surface westerlies of ≈ 2 m s^{-1}, this linear model predicts that the ratio of the orographic response to the thermal response (for $\kappa = 10$) increases by nearly a factor of 2.

4. Conclusions

The linear responses to orography and shallow heat sources in midlatitudes are sensitive to the magnitude of the mean low-level westerlies.

In the orographic case, the mean winds are small enough near the surface that meridional advection, $v'\partial\bar{T}/\partial y$, dominates zonal advection $\bar{u}\partial T'/\partial x$, and balances the adiabatic cooling due to the forced ascent over the topography, $N^2w' = N^2\bar{u}\partial h/\partial x$. As a result, the response is proportional to $\bar{u}$. While it seems intuitive that the orographic response should increase with the strength of the surface winds, note that the meridional temperature gradient is essential to this result.

In the case of extratropical heating, the term N^2w' tends to be small. If meridional advection were still dominant in balancing the prescribed heating, then the response would be more or less independent of $\bar{u}$. The key additional element in the response to heating is that a large part of the response at low levels has the scale of the heating rather than the generally larger scales of the vertically and horizontally radiating Rossby waves. This favors zonal advection (since $T' \propto \psi'_z$) when heating is shallow, and results in an increase in the response as $\bar{u}$ decreases.

Because the changes in the amplitudes of the orographically and thermally forced waves are of opposite sign, their relative proportion is particularly sensitive to the low-level mean flow. This sensitivity must be kept in mind when trying to use linear theory to determine the relative importance of orography and heating for the observed stationary waves.

The meridional displacement of a mean streamline near the surface, $\psi'/\bar{u}$, tends to be independent of $\bar{u}$ for orographic forcing, since $\psi' \propto \bar{u}$. For thermal forcing the meridional displacement tends to be $\propto \bar{u}^{-2}$. To the extent that thermal forcing dominates near the surface, the asymmetry of the low-level flow, according to linear theory, is exceptionally sensitive to the low-level mean winds.

The relevance of these results to the atmosphere remains to be demonstrated. It is natural to suppose that a model linearized about the zonal mean flow overestimates the sensitivity of the stationary eddies to changes in the mean flow, particularly when this flow is weak. Nonlinearity is likely to be especially important in the thermally forced case. For fixed heating, the linearly forced eddies grow as the mean flow weakens, so the zonal wind perturbations can easily become comparable to $\bar{u}$. In the orographic case, linear theory could clearly be misleading for a feature as large as the Tibetan plateau. The interaction between thermal and orographic forcing is also likely to be important (e.g., Chen and Trenberth 1988). Until a better understanding emerges of the range of validity of linear theory and the ways in which it breaks down, it is appropriate to reserve judgement.

One can also question the relevance of computing the sensitivity of the response (whether linear or nonlinear) to prescribed heating (e.g., Shutts 1987), since the heating field is in reality dependent on the mean flow and the stationary eddy response. If a parameter such as the strength of the surface drag is modified in a GCM so as to change the low level flow, there is no guarantee that the heating field would not change in such a way as to invalidate these scaling arguments.

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APPENDIX

A Problem with σ-Coordinate Models

The linear orographic response shown in Fig. 1 is computed using equally spaced levels in the vertical. When the computation is redone using the vertical levels of the GCM that served as the starting point for the construction of the linear model (σ = 0.025, 0.095, 0.205, 0.350, 0.515, 0.680, 0.830, 0.940, 0.990), one finds the result shown in Fig. 4. Although the response near $\lambda = 1$ is similar, the distortion is pronounced at small λ; the amplitude of the response when $\lambda = 0$ is still 30% of its amplitude at $\lambda = 1$.

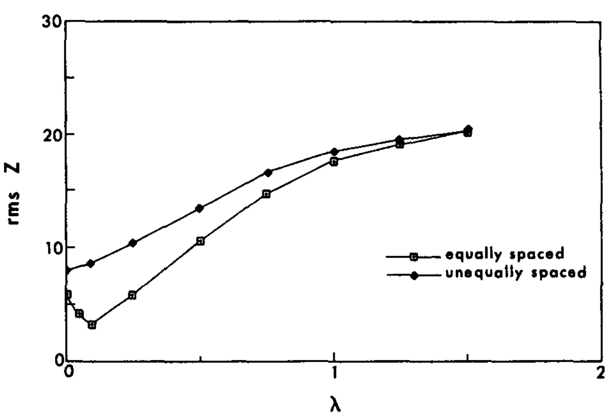

FIG. 4. Sensitivity of orographically forced stationary wave to zonal mean low level wind (as in Fig. 1) for σ-coordinate linear models with equally and unequally spaced vertical levels.

As the zonal mean surface wind tends to zero, the linear (inviscid) response to orography should also tend to zero. When using σ as vertical coordinate, this fact is disguised due to the relation between the σ-coordinate and pressure coordinate perturbations. For the zonal wind perturbations one has

$$u'_\sigma = u'_p - \sigma (\ln p_s)' \partial \bar{u}/\partial \sigma. \tag{A.1}$$

If there is orography, so that p'_s is nonzero, and if the flow is unperturbed in pressure coordinates (or z coordinates for that matter), as would be the case if the zonal mean surface flow were zero, the σ-coordinate perturbation is still nonzero as long as there is vertical shear in the basic state. (The difference between the zonal means in the two coordinate systems is of second order in the perturbations and can be ignored in the context of linear theory.) If one linearizes the inviscid σ-coordinate equations about a zonally symmetric state in which the mean surface wind is zero, one can demonstrate that these equations are satisfied by setting

$$u'_\sigma = -\sigma(\ln p_s)' \partial \bar{u}/\partial \sigma \tag{A.2}$$

$$T'_\sigma = -\sigma(\ln p_s)' \partial \bar{T}/\partial \sigma \tag{A.3}$$

$$\sigma' = -\sigma \bar{u} \partial (\ln p_s)'/\partial x \tag{A.4}$$

$$\Phi' = -RT(\ln p_s)', \tag{A.5}$$

consistent with the pressure coordinate disturbances being identically zero. But the cancellation of terms that is needed to obtain this result need not occur with sufficient accuracy in a finite-differenced model. As indicated in Fig. 4, the severity of this problem is evidently much greater in a model with the less accurate finite-differencing associated with unequally spaced levels.

The behavior seen for very small λ in the equally spaced case seems to be due to a different cause; it does not disappear as the vertical resolution is increased. We believe that the nonzero amplitude at $\lambda = 0$ in this case is primarily due to the frictional stresses. It is the σ-coordinate perturbation that is damped by the surface friction in our model, so the stress will be nonzero even when the pressure coordinate perturbation vanishes. As a result, a vanishing pressure coordinate perturbation is inconsistent. At least in the case of surface friction, it is physically appropriate that the damping have such an effect; it forces the wind back to zero at the top of the mountain, not to the mean flow at that pressure level.

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