# Linearized Calculations of Stationary Waves in the Atmosphere

By Richard S. Lindzen, Takehiko Aso[1] and David Jacqmin

*Center for Earth and Planetary Physics, Harvard University*
*Cambridge, Massachusetts 02138*
*(Manuscript received 12 August 1981)*

## Abstract

The linearized response of the atmosphere to thermal and orographic stationary wave forcing is examined with a high resolution (1 km vertical resolution, 1.5° latitude resolution) primitive equation model. It is found that the response to thermal forcing (but not orographic forcing) is sensitive to small changes in the basic distribution of wind and temperature. This suggests that anomalies in stationary waves can occur even without changes in forcing.

## 1. Introduction

The present paper discusses a few general considerations involved in modelling stationary waves in the atmosphere, and presents some preliminary, suggestive results.

In the past year, three other papers on this topic have appeared (Opsteegh and Van den Dool 1980, Hoskins and Karoly 1981, and Webster 1981). Each of these papers involved extremely coarse vertical resolution (2 layers for Opsteegh and Van den Dool and Webster, 5 levels for Hoskins and Karoly) they also restricted themselves to the response to a localized source of thermal forcing. The coarse resolution was justified with the claim that the atmospheric response was almost barotropic.

The purpose of the above studies was to examine the possibility that anomalous stationary wave patterns at middle and high latitudes could be produced by anomalous tropical heating. The purpose of the present paper is to see to what extent anomalous stationary waves can be produced by small changes in zonally averaged winds and temperatures—even when stationary forcing remains unchanged.

The present study uses a high vertical resolution model to show that barotropicity is closely linked to the localized nature of the forcing; the response to global forcing is significantly non-barotropic. Our study reveals important sensitivities to basic state variations, but the origin of these sensitivities is not entirely clear. Small variations in static stability in middle latitudes and in zonal winds near the equator are both involved. Tests have been performed to show that the results are not due to numerical errors.

## 2. Wave response model

Our primary model is basically that of Lindzen and Hong (1974) where instead of tidal forcing, we consider climatological distributions of stationary forcing derived from Ashe (1979) and Sankar-Rao (1965) for thermal and orographic forcing respectively.[1] In section 3 we will discuss this forcing in greater detail. We have also developed alternative models: i) A quasi-geostrophic model similar to Matsuno (1970); and ii) A primitive equation model where—in contrast to Lindzen and Hong—the equations are *not* reduced to a single unknown. The latter should be formally equivalent to our primary model and provides a convenient check for any results. In fact, differencing the original equations

---

1 Current Address: Radio Atmospheric Science Center, Kyoto University, Uji, Kyoto, Japan

1 In order to use the Lindzen-Hong model for the present problem we had to drop assumptions in Lindzen and Hong to the effect that contributions to static stability from the latitudinally varying part of the temperature are small, and that doppler shifts in frequency are small compared to damping rates and stationary frequency. The last is, of course, zero for stationary waves.

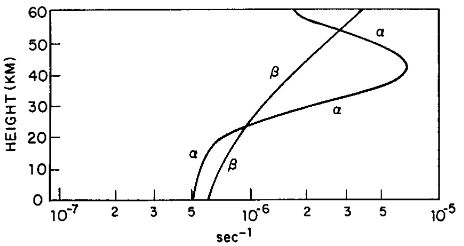



Fig. 1 Vertical distributions of Rayleigh friction and Newtonian cooling rate coefficients.

does involve a greater measure of numerical smoothing than occurs with the Lindzen and Hong model.

We separately consider zonal wavenumbers 1, 2, and 3, and instead of tidal periods we set frequency to zero. Figure 1 shows our vertical distributions of Rayleigh friction and Newtonian cooling rate. Above 25 km we use values larger than usually used in order to form an absorbing "sponge" region. When we use primitive equations, stationary gravity waves are excited in the neighborhood of the equator having very short vertical wavelengths (see Lindzen, 1967 for an analysis of the dispersive properties of such waves; Webster, 1972, encounters a similar problem). This condition leads to severe numerical problems in our primary model which we suppress by increasing Rayleigh friction by an order of magnitude in a 5° latitude neighborhood of the equator.
Interestingly, our alternative primitive equation model does not seem to need the added friction. Perhaps numerical smoothing plays the same role.

Our model has a resolution of 1 km in the vertical. We have 119 points between the poles. We have a top at 60 km. Enhanced damping above 25 km eliminates sensitivity to the presence of a lid. Also, increased resolution did not significantly change results.

## 3. Forcing models

*a) Topographic forcing*

Topography is introduced through the lower boundary condition

$$w=\frac{dz}{dt} \quad \text{at} \quad z=h(\theta, l) \qquad (1)$$

where

$w$ = vertical velocity
$z$ = height
$h$ = topographic elevation
$\theta$ = latitude, $l$ = longitude

We take stationary waves to be linearizable perturbations on a basic state consisting of a zonal wind in thermal wind balance. The conventional linearization of (1) is

$$w=\frac{U(\theta, z)}{a\cos\theta}\frac{\partial h}{\partial l} \quad \text{at } z=0 \qquad (2)$$

where

$U$ = basic zonal flow

It has commonly been noted (Dickinson, 1980) that even where the linearized equations of motion are adequate, the linearization of (1) may be inaccurate. Attempts to improve on this situation can, however, lead to certain inconsistencies, and we shall basically retain (2) except that we will, in some instances, evaluate $U$ at some height greater than $z=0$ (usually $z=2.5$ km). The use of the larger values of $U$ simply magnifies the topographic forcing and the resulting response by a fairly uniform 40%. The rationale for doing this is that $U$ changes significantly over the height of the topography and using the smallest value of $U$ (that for $z=0$) may underestimate topographic forcing. In point of fact results suggest that our use of $U$ at $z=2.5$ km probably overestimates the topographically forced stationary waves.

For $h$ we take the spherical harmonic expansion of topography given by Sankar-Rao (1965).

*b) Thermal forcing*

The notorious problems in ascertaining this quantity are reviewed by Dickinson (1980). Ashe (1979) has estimated the long term mean asymmetric component of sensible and latent heating over the northern hemisphere for July and January. No comparable estimates exist for the southern hemisphere. Moreover, it is not at all clear whether interannual variations in the asymmetric heating may not be as large as the long term mean. For the present, we will use Ashe's model. We smoothly extend Ashe's northern hemisphere results to the southern hemisphere tropics—but let thermal forcing go to zero beyond 30°S.[2] As noted by Dickinson (1980),

2 For southern hemisphere thermal forcing we take northern hemisphere forcing shifted by six months and weighted by:
$\exp[-(\theta-90°/28°)^2]$, where $\theta$ is colatitude in degrees.
This provides a smooth global forcing though the specific choice has no observational basis.

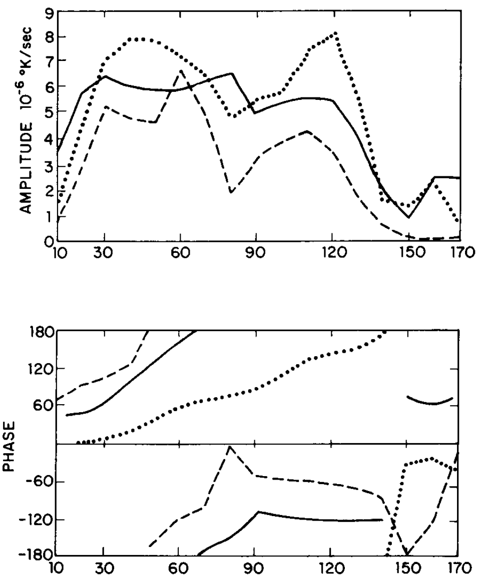



Fig. 2 Latitude variation of amplitude and phase of vertically integrated thermal forcing in zonal wavenumbers 1, ——— 2, ......, 3. – – – –.

Ashe's results are uncertain by as much as a factor of two.

It is our impression that our present model hardly transmits planetary waves across the equator, so that the extension of our forcing across the equator is of little consequence for northern hemisphere results. Figure 2 shows Ashe's results decomposed into wavenumbers 1, 2, and 3. It is interesting to note that in winter, thermal forcing is not concentrated in the tropics. Also, the phase of thermal forcing changes rather markedly with latitude.

We next turn to the vertical distribution of heating. Ashe assumed all heat was deposited below 500 mb. This choice was motivated by his use of a 2-layer model. However, this is inappropriate for the tropics. In fact the choice of vertical distribution in the tropics is relatively simple. We choose, for $z \leq z_T$,

$$J = j(\theta) e^{z/H} \sin\left[\pi\left(1 - \frac{z}{z_T(\theta)}\right)\right] \qquad (3)$$

where

$H \simeq$ scale height $\simeq 8$ km.

Above $z_T$, $J$ is set equal to zero. $z_T$ is taken to be that height at which the atmosphere's dry enthalpy equals the moist enthalpy of surface air: *i.e.*,

$$\begin{aligned} C_p T(0) + Lq(0) &= C_p T(z_T) + g z_T \\ &\approx C_p\{T(0) - \Gamma z_T\} + g z_T \qquad (4) \end{aligned}$$

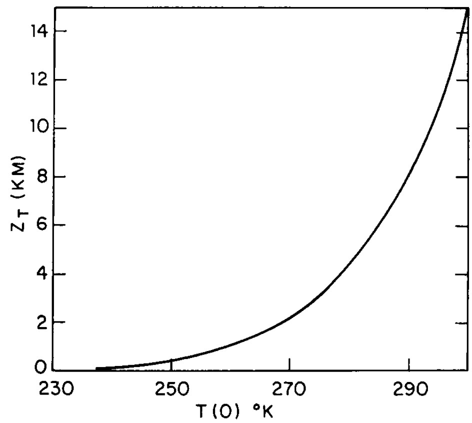



Fig. 3 Estimated height of convective layer as a function of surface temperature.

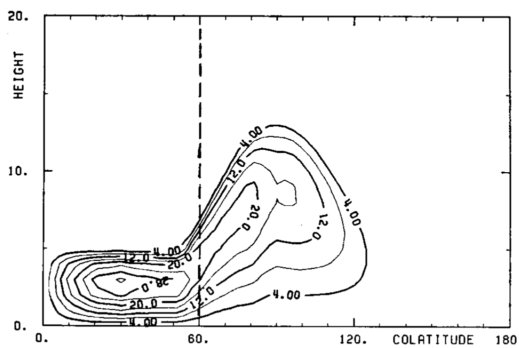



Fig. 4 Meridional cross-section of wavenumber one thermal forcing, in milliwatts per kilogram.

Here $L$ is the heat of condensation, $q(0)$ is the mixing ratio of water vapor at ground level, and $\Gamma$ is the adiabatic lapse rate.

For $q(0)$ we assume 75% humidity at $T(0)$.[3] Figure 3 shows $z_T$ as a function of $T(0)$ based on (4).

Equation (4) is adequate in the tropics where the temperature variance is small. However, in middle latitudes the zonally averaged $T(0)$ implies very small values for $z_T$. The actual $z_T$ is largely determined by the temperature variance. The inclusion of this feature seems rather complicated; instead we have simply taken $z_T = 5$ km whenever (4) implies a smaller value.

Figure 4 shows a meridional cross-section of the amplitude of forcing for zonal wavenumber 1 (other wavenumbers are similar). Note that in

3 This approach is based on the cumulus parameterization of Lindzen (1981). A simplified version of this parameterization is given in Stevens and Lindzen (1978).

the tropics thermal forcing peaks above 500 mb. Note that $j(\theta)$ in (3) is determined so that

$$\int_0^\infty J\, e_0 dz = \text{Ashe's heating.}$$

In section 6 we will discuss tropical thermodynamics further in connection with stationary wave responses to sea surface temperature anomalies.

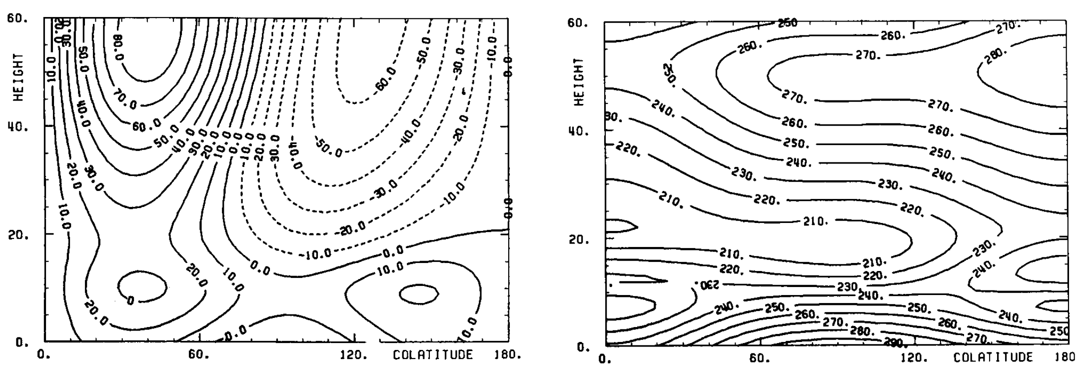
## 4. Models for zonally averaged winds and temperatures

Basic zonal wind and temperature are assumed to be in thermal wind balance. This is guaranteed by the analytic specification of Lindzen and Hong (1974; see p. 1933 J.A.S. *31* for erratum). Although the present calculations use this truncated analytic specification, it has some important difficulties: most notably, the truncation of the representation makes it difficult to modify the basic wind and temperature in one height-latitude region without modifying them elsewhere. Our basic intention was to investigate the sensitivity of stationary wave responses to whether zonal winds in the tropical upper troposphere were westerly or easterly. Such changes, it was found, inevitably altered the midlatitude jets—until the position of the midlatitude jets was moved somewhat poleward of their climatological positions.

The two wind and temperature distributions we finally adopted are shown in Figures 5 and 6. Clearly, these distributions differ primarily in the tropics, and as we shall soon see, the stationary wave responses for these two distributions are significantly different. We are, however, still uncertain as to why the responses are different, and current calculations suggest that it may be important that our two distributions differ not only with respect to tropical winds but also changes in midlatitude lower tropospheric static stabilities. The latter vary by as much as 1°/km at 5 km—but by less elsewhere.

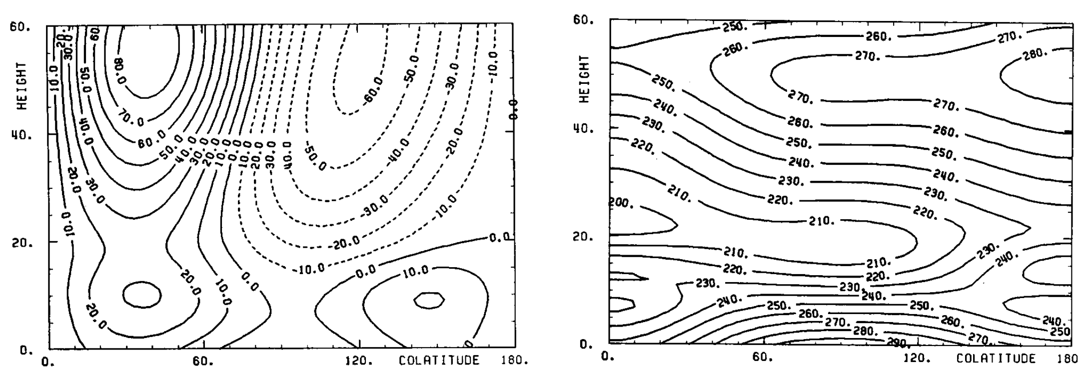
## 5. Some results

One obvious difficulty with studies of the present type is the shear volume of results. Little possibility exists for a complete presentation. Also, given our present uncertainty concerning




(a) (b)

Fig. 5. Meridional contours of basic zonal wind (m/s) and temperature (°K).




(a) (b)

Fig. 6. Meridional contours of basic zonal wind (m/s) and temperature (°K).

the interpretation of our results, there seems little justification for an elaborate presentation. Nevertheless, the examination of a few selected results may prove illuminating.

*a) Topographic forcing*

In Figures 7, 8, 9 we show the amplitude and phase of the response in geopotential height to topographic forcing for wavenumbers 1, 2 and 3 respectively. The basic state shown in Figure 6 is used, and equation (2) was used with $U$ taken from $z=2.5$ km (amplitudes would be 30% less if we used $z=0$ km instead). Results are shown for northern hemisphere winter. Several aspects of these results warrant comment.

i) The results for the troposphere are remarkably similar in magnitude, phase, and structural features to van Loon, *et al*'s (1973) average observed, zonal waves. To be sure there is some shift in latitude between observed and predicted details—but this is to be expected in view of the fact that features in Figure 6 are also displaced from their observed positions. Also our amplitudes are a little larger—probably because of our lower boundary condition. What is perhaps most remarkable is that there should be such good agreement between observations and calculations based on topographic forcing alone. This was already suggested in the work of Charney and Eliassen (1949).

As we shall soon see, our response to thermal forcing is indeed significantly smaller. However, given the uncertainty in Ashe's forcing, this finding may turn out to be incorrect.

ii) Our predicted amplitudes in the stratosphere

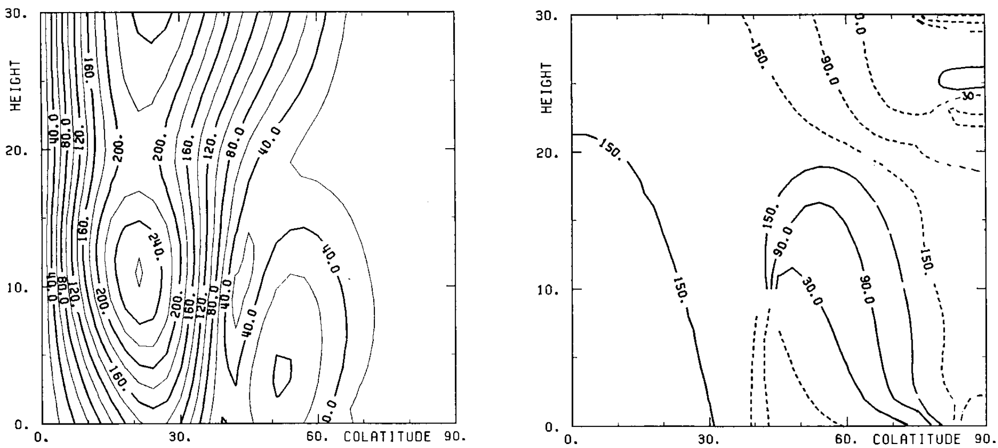



Fig. 7 Meridional contours of amplitude and phase for the geopotential height response to wavenumber 1 orographic forcing (basic fields shown in Fig. 6 were used).

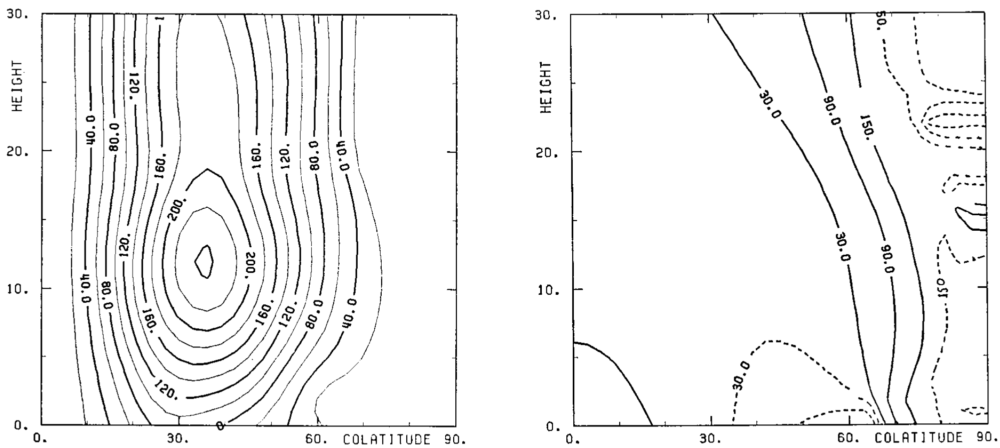



Fig. 8 Same as Fig. 7 but for wavenumber 2.

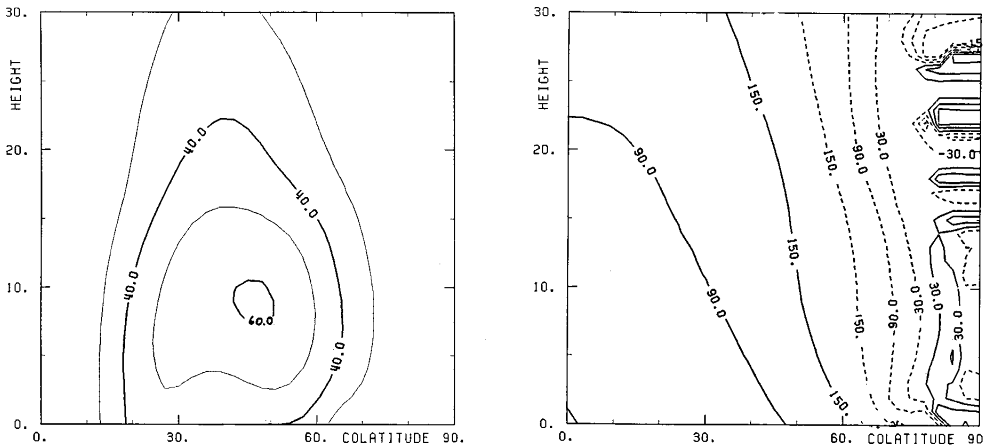



Fig. 9 Same as Fig. 7 but for wavenumber 3.

are weaker than observed—largely due to our enhanced damping.

iii) The presence of short vertical wavelengths over the equator is evident at all zonal wavenumbers.

iv) In neither the calculations nor the observations is there any evidence of a barotropic response (except possibly for $s=3$ where we have only marginal vertical propagation).

*b) Response to thermal forcing*

Figure 10 shows the amplitude and phase of the geopotential height field associated with the $s=1$ response to thermal forcing for northern hemisphere winter conditions. Again the basic state from Figure 6 is used. Figure 4 shows a rough breakdown of how forcing is divided into tropical and higher latitude components. Figures 11 and 12 show the specific $s=1$ response to these two components of the forcing. Note that the response to high latitude forcing appears to be larger. Moreover, the responses to tropical and midlatitude forcing are sufficiently different so that the total response is generally smaller than the response to high latitude forcing alone. As already noted, the total response to thermal forcing is appreciably weaker than the response to topography. However, as long as the possibilities remain that thermal forcing may be a factor of two larger and orographic forcing 30% smaller, it is difficult to be sure that thermal and orographic responses might not be comparable. We will also see, in section 5, that responses to thermal forcing are larger for the basic state shown in Fig. 5.

Note also that the response at higher latitudes to higher latitude forcing is quite baroclinic, but

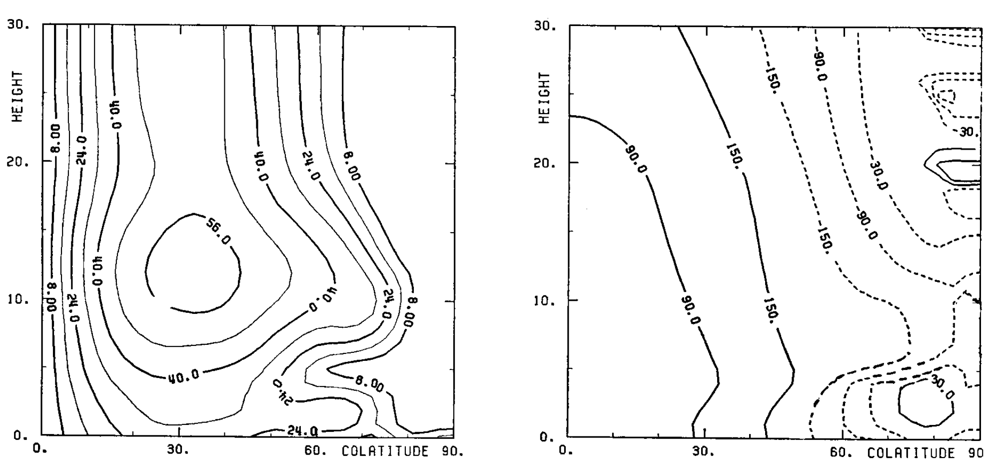



Fig. 10 Same as Fig. 7 but for wavenumber 1 response to global thermal forcing.

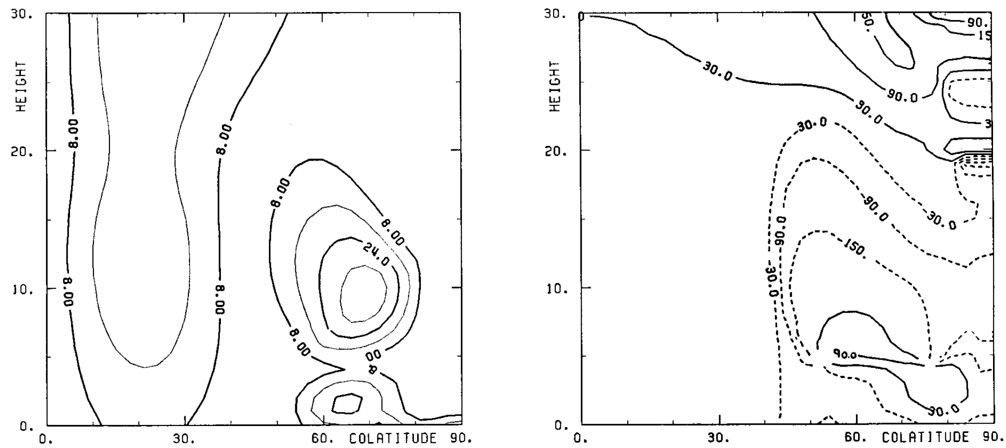



Fig. 11 Same as Fig. 7 but for wavenumber 1 response to tropical component of thermal forcing.

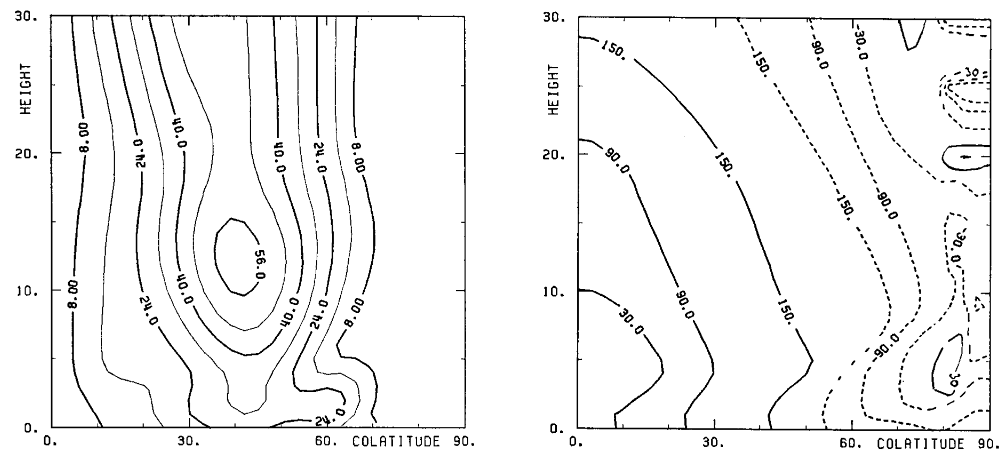



Fig. 12 Same as Fig. 7 but for wavenumber 1 response to middle and high latitude component of thermal forcing.

the response to tropical forcing is reasonably barotropic. The reason is quite clear. In propagating from low to high latitudes, baroclinic components disperse vertically leaving primarily the barotropic component at great distances. Such an effect is clearly irrelevant to local forcing and certainly does not support the assumption of barotropicity as a general rule.

Figures 13, 14, and 15 show results for $s=2$. In this case there is significant cancellation between components forced in high and low latitudes. Responses at $s=3$ were small and are not shown.

*c) Sensitivity to basic state*

Figures 16, 17 and 18 reproduce the results shown in Figures 10, 11 and 12—but for the basic state shown in Figure 5. While gross structures remain similar, there are numerous quantitative differences. Some seem superficially reasonable. For example, Figure 17 shows an enhanced response to tropical forcing. Now Figure 5 shows unbroken westerlies across the tropical upper troposphere while Figure 6 shows easterlies in the tropical upper troposphere. Presumably, the westerlies permit the transmission of stationary waves thermally forced in the tropics where thermal forcing extends into the upper troposphere. The question, however, remains as to why in Figure 18, we also see enhanced response to midlatitude forcing.

We will discuss the above results in the next section. First, however, we wish to present some additional results for wavenumbers 2 and 3, and for topographically forced stationary waves.

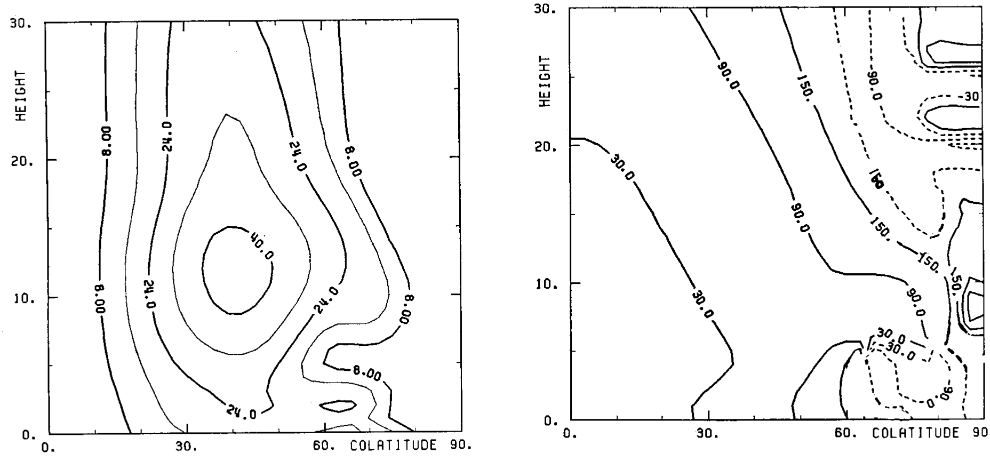



Fig. 13 Same as Fig. 10 but for S=2.

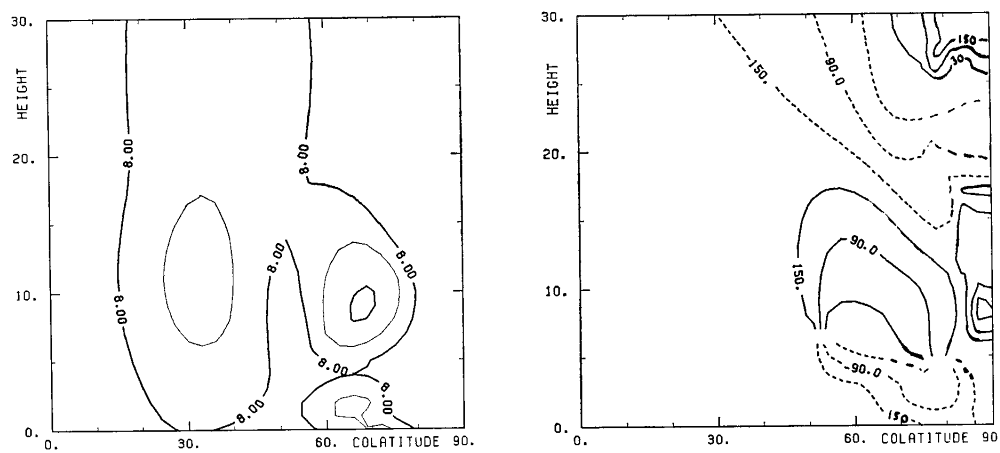



Fig. 14 Same as Fig. 11 but for S=2.

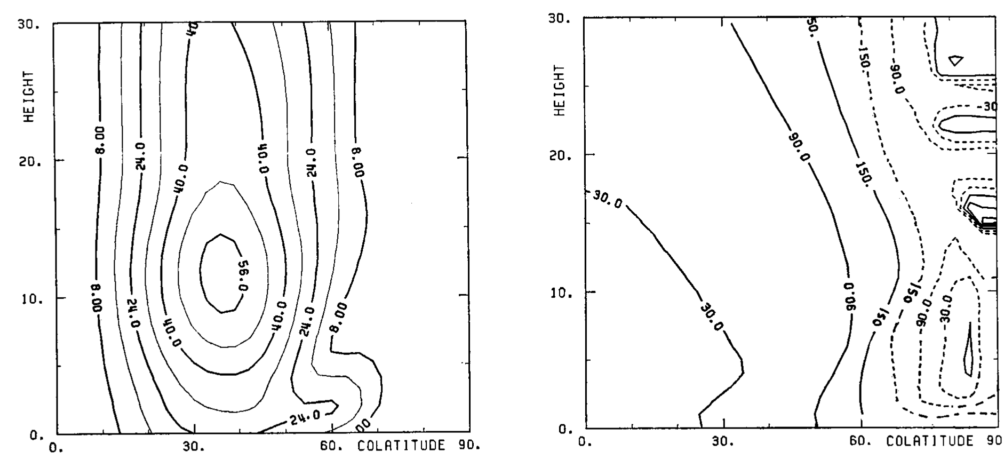



Fig. 15 Same as Fig. 12 but for S=2.

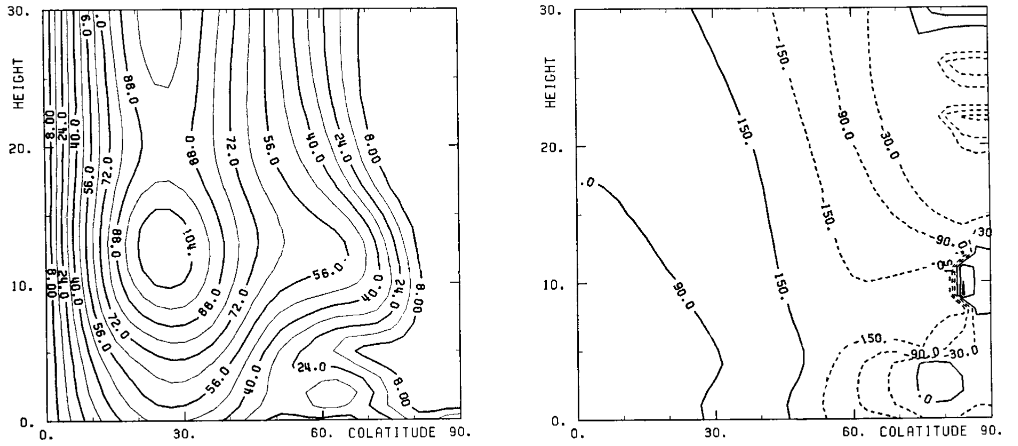


Fig. 16 Same as Fig. 10 but for basic state shown in Fig. 5.

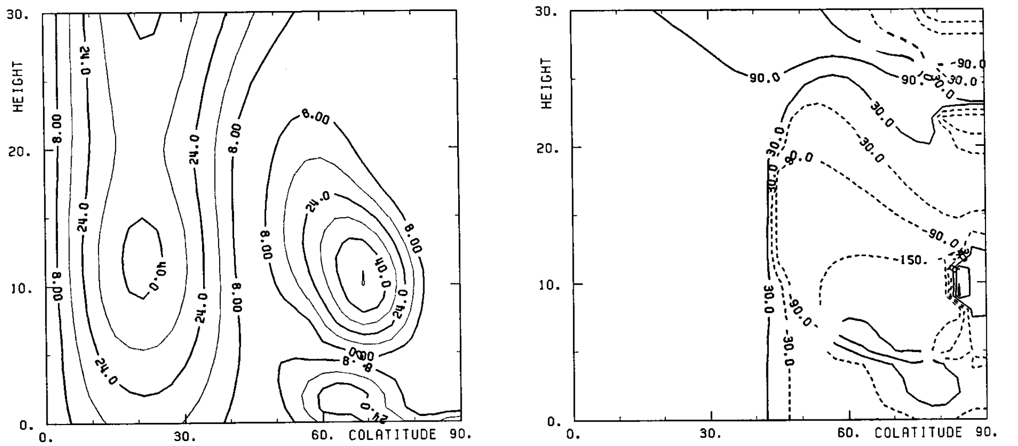


Fig. 17 Same as Fig. 11 but for basic state shown in Fig. 5.

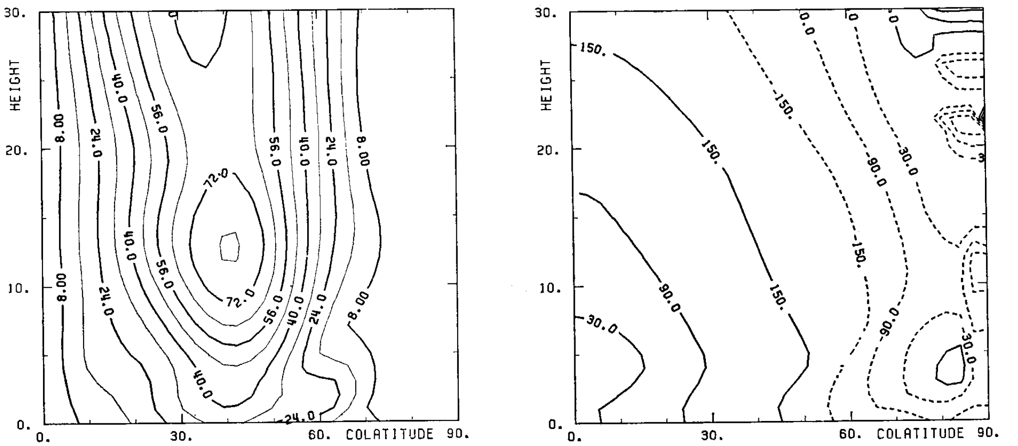


Fig. 18 Same as Fig. 12 but for basic state shown in Fig. 5.

Table 1 Geopotential Height at 5 km.

| | | | 60° | | | 30° | | |
|---|---|---|---|---|---|---|---|---|
| Zonal # | Forcing | Basic State | Amp | Phase (° Long) | Ridge Position (° Long) | Amp | Phase (° Long) | Ridge Position (° Long) |
| 1 | TH | 5 | 18 | −144 | 144 | 74 | 90 | −90 |
| | | 6 | 12 | −150 | 150 | 42 | 85 | −85 |
| | TR | 5 | 6 | 30 | −30 | 25 | 45 | −45 |
| | | 6 | 10 | 90 | −90 | 13 | −12 | +12 |
| | HL | 5 | 16 | −150 | +150 | 38 | 94 | −94 |
| | | 6 | 12 | −150 | 150 | 25 | 90 | −90 |
| | TO | 5 | 65 | 26 | −26 | 174 | 150 | −150 |
| | | 6 | 61 | 32 | −32 | 162 | 155 | −155 |
| 2 | TH | 5 | 16 | 127 | −64 | 38 | −13 | 6 |
| | | 6 | 8 | 130 | −65 | 25 | 28 | −14 |
| | TR | 5 | 5 | −76 | 38 | 25 | −113 | 57 |
| | | 6 | 8 | −40 | 20 | 15 | −170 | 85 |
| | HL | 5 | 17 | 127 | −64 | 50 | 24 | −12 |
| | | 6 | 12 | 142 | −71 | 39 | 28 | −14 |
| | TO | 5 | 49 | 3 | −1 | 189 | −16 | 8 |
| | | 6 | 48 | 4 | −2 | 162 | −16 | 8 |
| 3 | TH | 5 | 3 | −176 | 59 | 15 | 32 | −11 |
| | | 6 | 7 | −163 | 54 | 7 | 15 | −5 |
| | TR | 5 | 3 | 25 | −8 | 8 | −23 | 8 |
| | | 6 | 2 | 166 | −55 | 6 | −72 | 24 |
| | HL | 5 | 7 | −170 | 57 | 11 | 64 | −21 |
| | | 6 | 7 | −167 | 52 | 6 | 63 | −21 |
| | TO | 5 | 55 | 161 | −54 | 81 | 75 | −25 |
| | | 6 | 47 | −175 | +58 | 54 | 75 | −25 |

Table legend: Response at 5 km and 60° and 30° colatitudes of basic states shown in Figs. 5 and 6 to: TR, tropical heating; HL, high latitude heating; TH, tropical and high latitude heating; TO, topography.

Rather than present detailed results for all these cases, we will instead concentrate on amplitudes and phases for $z=5$ km, and both 30° latitude and 60° latitude. Table 1 shows these results for both basic wind profiles, the various choices of thermal and topographic forcing and wavenumbers 1, 2 and 3. We see that the wavenumber 3 response to thermal forcing is rather weak. The wavenumber 2 response is sizeable and very sensitive to the choice of basic state. For the most part, the response to topographic forcing is relatively insensitive to basic state, although at $s=3$ the amplitude of the response is fairly sensitive.

## 6. Some remarks on calculated sensitivities

In section 5 we showed that the atmosphere's response to thermal forcing depends very significantly on relatively small variations in the zonally averaged fields. The differences shown in Table 1 represent major changes in the position of large scale features and presumably in storm paths (Niehaus, 1980). To the extent that these changes resulted from changes in tropical winds, we have some potentially serious problems for midlatitude weather forecasting, since tropical winds are not very well measured. The general wisdom appears to be that changes in stationary waves arise from changes in sea surface temperature and the resulting changes in forcing (Wallace & Gutzler 1981). The changes in forcing arise from changes in evaporation (which can amount to 20% in the tropics)—and, more importantly, from the fact that perturbations in surface temperature can induce major modifications in low level moisture convergence (Schneider & Lindzen, 1977, Schneider, 1977). The fact that both these processes favor the tropics implies a major role for the tropics in producing anomalies in stationary waves. Nonetheless, the present work suggests that significant anomalies can be produced without changes in forcing.

This result, while potentially important, requires critical assessment. Even the question of whether it is meaningful to assess the linearized response of a fixed basic state to specified forcing is unclear. The presence of damping (especially at surfaces of zero wind) implies that the basic state is changing. The present calculations assume that this is occurring slowly, and that zero wind lines are absorbing. Even without questioning the basic utility of our approach, there remain two questions which must be addressed:

i) Are the results we obtain, numerically correct solutions of the basic equations, and

ii) What is the physical basis for our results?

In order to answer (i) we did a number of conventional checks. Modest changes in resolution did not produce significant changes in results. Apart from this, we attempted to duplicate the results of our primary model using two different models:

i) A quasi-geostrophic model similar to Matsuno (1970), and

ii) A primitive equation model where, instead of solving a single equation for geopotential, one starts with the basic equations for momentum, mass and energy, and solves the whole array.

We will present a detailed description of these computations in the future. However, the following results were found:

i) The sensitivities shown in Table 1 were much reduced in the quasi-geostrophic model, and

ii) The primitive equation array closely duplicated the results of Table 1, for $s=1$. Other wave numbers are currently being checked.

While result (ii) lends confidence to our numerical results, the difference between results (i) and (ii) are perplexing. In particular, our simplistic notions about the role of zero wind lines in permitting thermally forced waves in the tropics to penetrate to middle latitudes should be applicable to both quasi-geostrophic and primitive equation models. Nevertheless, the sensitivities in both cases were different. We were, therefore, forced to consider the various ways in which the two models differed. One way has to do with static stability: quasi-geostrophic models use a fixed, globally averaged vertical distribution of static stability, while primitive equation models allow static stability to vary. This led us, in turn, to see whether the basic states shown in Figures 5 and 6 differed significantly with respect to static stability. To our surprise we found that static stabilities at middle latitudes did vary by about 1°/km near $z=5$ km. We next performed a somewhat artificial numerical experiment wherein we examined the response of a primitive equation model for the wind distributions shown in Figures 5 and 6, but where the temperature field from Figure 6 was employed in both cases. The results were fairly close to those found using the quasi-geostrophic model. These tests suggest a rather simple—but important—difference between quasi-geostrophic and primitive equation models. It is self-evident that distributions of zonally averaged winds do change and that such changes must be accompanied by modest changes in static stability. The present results suggest that these changes in static stability can significantly alter stationary wave responses.

## 7. Concluding remarks

Our current research on this problem is in a state of substantial uncertainty. The sensitivity of stationary wave response to changes in static stability was unanticipated. Nevertheless, sensitivity to either static stability or wind of the order shown in Table 1, suggests that marked changes in stationary waves can occur even in the absence of significant changes in forcing. It remains for future work to disentangle the various causes and effects in our preliminar studies and to elucidate the relevant physical mechanisms. Certainly, the shortcomings of our analytic representation of the basic state have become clearer. In order to better determine the nature of stationary wave sensitivities, we will have to change to a specification of the basic state wherein changes can be restricted to specified regions. In such studies we will, of course, study sensitivities not only to tropical winds and low level static stabilities. Sensitivities to winds and temperatures at other latitudes and levels will also be considered.

## References

Ashe, S., 1979: A nonlinear model of the time-average axially asymmetric flow induced by topography and diabatic heating. *J. Atmos. Sci.*, **36**, 109-126.

Charney, J. G. and A. Eliassen, 1949: A numerical method for predicting the perturbations of the mid-latitude westerlies. *Tellus*, **1**, 38-54.

Dickinson, R. E., 1980: Planetary waves: theory and observation in *Orographic Effects in Planetary Flows, GARP Publication Series*, **23**, World Meteorological Organization.

Hoskins, Brian J. and Karoly, David J., 1981: The steady linear response of a spherical atmosphere to thermal and orographic forcing. *J. Atmos*

*Sci.*, **38**, 1179-1196.
Lindzen, R. S., 1967: Planetary waves on beta planes, *Mon. Wea. Rev.*, **95**, 441-451.
Lindzen, R. S., 1981: Some remarks on cumulus parameterization. Proceedings of the NASA Cloud, Climate Conference, NASA Report, Available NASA/Goddard Institute of Space Studies.
Lindzen, R. S. and S. S. Hong, 1974: Effects of mean winds and horizontal temperature gradients on solar and lunar semi-diurnal tides in the atmosphere. *J. Atmos. Sci.*, **31**, 1421-1446.
Matsuno, T., 1970: Vertical propagation of stationary planetary waves in the winter northern hemisphere. *J. Atmos. Sci.*, **27**, 871-883.
Niehaus, M. C. W., 1980: Instability of non-zonal baroclinic flows. *J. Atmos. Sci.*, **37**, 1447-1463.
Obsteegh, J. D. and Van den Dool, H. M., 1980: Seasonal differences in the stationary response of a linearized primitive equation model: Prospects for long-range weather forecasting? *J. Atmos. Sci.*, **37**, 2169-2185.
Sankar-Rao, M., 1965: Continental elevation influence on the stationary harmonics of the atmospheric motion. *Pure Appl. Geophys.*, **60**, 141-159.
Schneider, E. K., and R. S. Lindzen, 1977: Axially symmetric steady-state models of the basic state for instability and climate studies. Part I: Linear calculations. *J. Atmos. Sci.*, **34**, 263-279.
Schneider, E. K., 1977: Axially symmetric steady-state models of the basic state for instability and climate studies. Part II: Nonlinear calculations. *J. Atmos. Sci.*, **34**, 280-296.
Stevens, D. and R. S. Lindzen, 1978: Tropical wave CISK with a moisture budget and cumulus friction. *J. Atmos. Sci.*, **35**, 940-961.
Van Loon, H., R. L. Jenne and K. Labitzke, 1973: Zonal harmonic standing waves. *J. Geophys. Res.*, **78**, 4463-4471.
Wallace, J. M., and D. S. Gutzler, 1981: Teleconnections in the geopotential height field during the northern hemisphere winter. *Mon. Wea. Rev.*, **109**, 784-812.
Webster, P. J., 1972: Response of tropical atmosphere to local, steady forcing. *Mon. Wea. Rev.*, **100**, 518-541.
———, 1981: Mechanisms determining the atmospheric response to sea surface temperature anomalies. *J. Atmos. Sci.*, **38**, 554-571.

## 線形論による大気中の定常波の計算

**Richard S. Lindzen, 麻生武彦*, David Jacqmin**
Center for Earth and Planetary Physics, Harvard University

高分解能（鉛直 1 km, 緯度方向約 1.5 度）のプリミティブ方程式モデルによって，定常な地形および熱的強制に対する大気の線形応答を調べた。熱的強制によって生じる定常波は基本状態の風や温度分布の小さい変動に対して敏感である。しかし地形性強制に対してはそうではない。この事実は定常波のアノマリーが，強制源の変化がたとえなくとも起こり得ることを示唆している。

*現所属：京都大学，超高層電波研究センター（京都府宇治市五ケ庄）