PROSPECTS FOR TROPICAL MODELING

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1. INTRODUCTION

A priori, the modeling of the tropics might appear to present problems no different from those found at middle latitudes. To be sure, cumulus convection is especially important in the tropics, but it is also important in middle latitudes. As it turns out, however, there are indeed special problems in modeling the tropics that, for the most part, pertain to the smallness of the coriolis parameter. In this lecture, I will discuss three aspects of tropical modeling covered in various papers. First I will consider the issue referred to in Lindzen and Fox-Rabinovitz (1989) (LF) as consistent resolution. Next, I will discuss the importance of circulations forced by surface temperature gradients, and the dependence of these circulations on the proper specification of the trade wind boundary layer. This section follows Lindzen and Nigam (1987) (LN). Finally, I will discuss the remarkable dependence of the ability of easterly waves to organize convection on almost unmeasurable details of the wind structure. This last item is part of the Ph.D. thesis of Miller (1990).

2. CONSISTENT RESOLUTION

LF suggested that there should be a balance between horizontal and vertical resolution in both numerical models and in observing systems. Intuitively, this is reasonable: smaller horizontal scales are, under some circumstances, associated with smaller vertical scales (and *vice versa*). If one resolves one scale but not the other obvious difficulties can arise. In the case of hyperbolic systems this leads to the well known Courant-Lewy-Friedrichs instability. However, for elliptic systems (which is more nearly the way our models behave in the spatial dimensions), the situation is less clear. The smaller scales tend to be excited only at boundaries or in the presence of forcing (wave-mean flow interaction, for example).

For quasi-geostrophic flows, the consistency relation between horizontal and vertical resolution is given by

$$\Delta L = \frac{N}{f_0} \Delta z \qquad (1)$$

while, for internal gravity wave type systems, the relation is

$$\Delta z = \frac{\sigma_i}{N} \Delta L, \qquad (2)$$

where f_0 is the local coriolis parameter, N is the Brunt-Vaisala frequency, $(1/\sigma_i)$ is the damping time, ΔL is the horizontal grid spacing, and Δz is the vertical spacing. What was not entirely clear in LF was how inconsistent resolution would manifest itself. To some extent this would depend on the initial conditions. To avoid such ambiguities we devised the following test: We took a limited area model with a physical domain of 4000 km by 4000 km and periodic boundary conditions. We ran this model with a range of vertical and horizontal resolution in various combinations. Our initial state consisted in a single sine wave on which random noise was superposed. We examined how long it took this model to blow up as a function of resolution. What we found is shown in Table 1. There is a reasonably clear tendency for the model to run longer when the resolution lies along the diagonal (i.e., when we have consistent resolution).

To evaluate the role of consistent resolution in data systems we did the following: We used the GLA model to perform analyses based only on satellite data over the oceans. The analyses were then filtered over increasingly broad horizontal scales. At each degree of filtration, the analyzed data was verified against available ground based radiosonde data. The RMS errors for the verifications are shown in Table 2 and Figure 1. At most levels the minimum verification error is achieved for nearly consistent resolution. The 250 mb and 300 mb levels are an exception; for these levels the lowest RMS error was obtained for the unfiltered analysis (perhaps supporting the contention that these are the equivalent barotropic levels). When we restrict our study to tropical latitudes, we find a stronger tendency for verification errors to minimize for large amounts of filtration (viz

MODEL LIFETIME ESTIMATES (in days) FOR DIFFERENT COMBINATIONS OF HORIZONTAL AND VERTICAL RESOLUTION

No. of Levels	Delta L (km) / Delta z (km)	400	200	100	50
5	3	17	11.5	8.5	6.5
10	1.5	9.5	12.5	8.5	7.5
20	0.75	7	15	9.5	10.5
40	0.375	6	8.5	8.5	13.5

(Maximum lifetimes highlighted)

Table 1. Model lifetime estimates (in days) for different combinations of horizontal and vertical resolution. See text for details.

VERIFICATION OF SATELLITE BASED DATA ANALYSIS OVER THE OCEANS V. SMOOTHING

(Minimum verification error highlighted)

LEVEL	SMOOTHING						
	1	2	3	4	6	8	10
50	82.13	82.75	82.79	84.34	84.54	82.34	81.95
70	78.85	79.46	79.41	79.50	79.38	77.82	78.03
100	65.33	65.69	65.84	65.85	65.26	64.30	64.36
150	49.32	49.45	49.35	49.82	48.52	49.43	53.34
200	37.66	37.81	37.26	37.96	37.29	42.20	53.13
250	31.81	32.10	32.21	32.77	34.15	41.02	59.53
300	31.79	32.09	31.85	31.93	36.78	49.69	70.13
400	28.86	29.07	28.78	28.07	33.46	44.19	64.58
500	25.96	25.85	25.24	23.88	24.71	34.90	48.25
700	25.61	25.30	25.33	24.96	25.05	32.51	39.79
850	23.78	23.60	23.26	23.26	23.63	30.05	33.93
1000	19.00	18.15	20.14	19.71	20.25	16.75	25.86

Table 2. RMS verification errors (in meters) for analyzed satellite data over the oceans subject to varying degrees of smoothing. Smoothing is over a model grid interval, so that a smoothing of 2 corresponds to two grid intervals, etc. The grid interval was approximately 250 km.

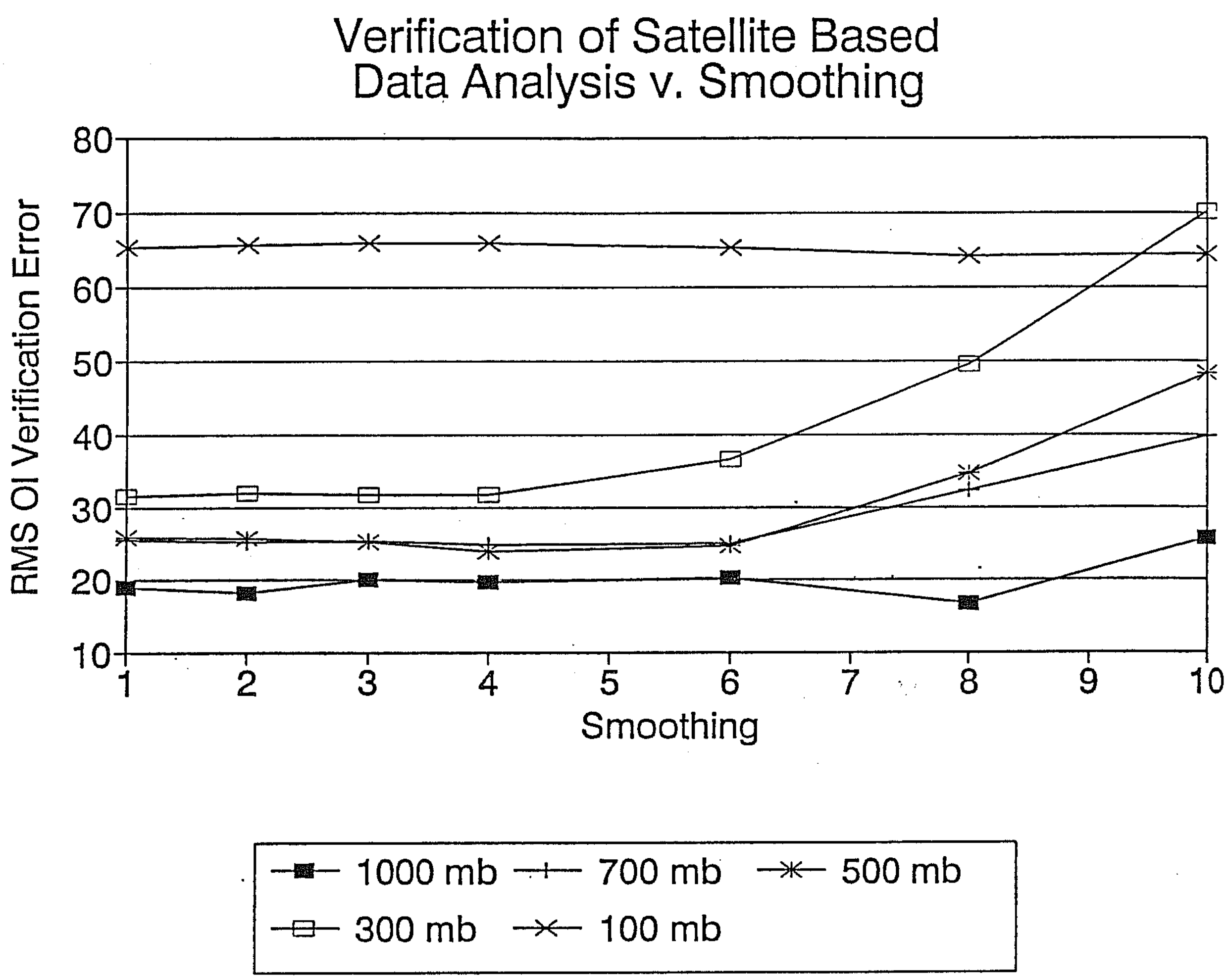

Figure 1. RMS verification errors (in meters) for analyzed satellite data over the oceans as a function of smoothing at various levels. Smoothing is over a model grid interval, so that a smoothing of 2 corresponds to two grid intervals, etc. The grid interval was approximately 250 km.

Table 3 and Figure 2). Even casual scrutiny of Figures 1 and 2 shows, however, that the variation of RMS verification error with smoothing is actually pretty small until the smoothing exceeds what is needed for consistent resolution. This, itself is significant. It says that no advantage is gained from horizontal resolution beyond that required for consistency with the fixed vertical resolution. As we will note in the next section, there is a genuine need for fine horizontal resolution. What the present results suggest is that the advantages of higher horizontal resolution cannot be fully achieved without consistent vertical resolution. One particularly disturbing feature emerges from a comparison of Tables 2 and 3. Normally, one might expect verification errors in geopotential height (which is what is listed in Tables 2 and 3) to be smaller in the tropics because geopotential variations in the tropics are small. This is indeed what is found for the most part below 400 mb; however, at upper levels (especially at 70 mb and 50 mb) verification errors in the tropics are actually larger.

What does the above tell us about the tropics? Clearly, if consistent resolution is meaningful, it places much greater demands on vertical resolution in the tropics. Equation 1 tells us that the needed vertical scale for quasigeostrophic motions is proportional to f; $f \rightarrow 0$ in the tropics; similarly, Equation 2 (as shown in LF) suggests that the possibility of critical surfaces introduces great demands for vertical resolution for gravity waves, and gravity waves appear to be a more prominent part of the tropical circulation.

It has frequently been pointed out that although motions on the finest resolvable horizontal scale may call for high vertical resolution, most important motion systems have much larger horizontal scales, and hence, don't demand such high resolution. There is some truth to this assertion. In the above described numerical experiment, had we not added random noise to our initial condition, the model would have run much longer -- regardless of resolution -- although runs with consistent resolution would still have run the longest before blowing up (according to the few cases we ran). The point is that models will generate all scales whether they are present in the initial conditions or not; however, the generation of the finest resolvable scales may take a long time if these scales are not present in the initial conditions. Thus, it may be possible to make useful

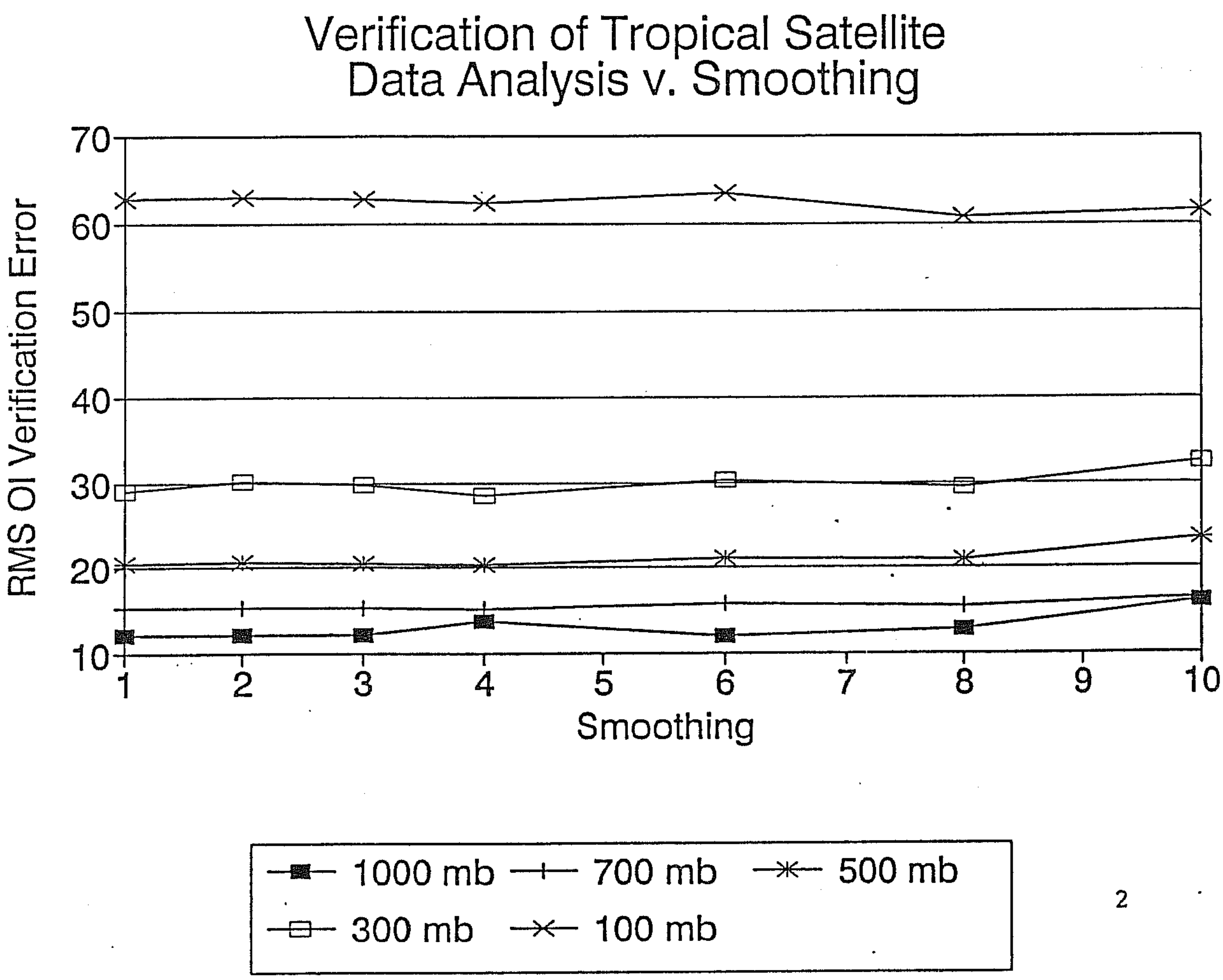

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Figure 2. The same as Table 2, but restricted to latitudes equatorwards of $\pm 30°$.

VERIFICATION OF SATELLITE BASED DATA ANALYSIS
OVER THE TROPICAL OCEANS V. SMOOTHING
(Minimum verification error highlighted)

LEVEL	SMOOTHING						
	1	2	3	4	6	8	10
50	105.45	105.56	105.84	106.11	104.33	102.96	108.38
70	112.31	112.27	112.29	112.73	112.41	111.29	113.23
100	62.81	63.03	62.86	62.37	63.44	60.80	61.55
150	41.53	41.76	41.30	41.49	41.91	41.68	41.68
200	41.86	41.88	41.52	40.68	40.28	39.49	42.46
250	36.36	36.58	36.14	35.60	35.24	34.36	37.06
300	29.00	30.23	29.89	28.56	30.35	29.48	32.62
400	25.96	26.20	25.81	26.62	26.16	25.58	28.01
500	20.57	20.75	20.52	20.42	21.23	20.96	23.59
700	15.29	15.33	15.32	15.21	15.76	15.45	16.48
850	14.50	14.56	14.25	14.13	14.39	13.18	15.58
1000	12.12	12.19	12.21	13.73	12.00	12.70	16.14

Table 3. The same as Table 2, but restricted to latitudes equatorwards of ±30°.

forecasts over finite times even if resolution is not consistent.

At the same time, the inability of present models to produce such features as the quasi-biennial oscillation suggests that there are real penalties incurred by inadequate vertical resolution.

3. FORCING BY SURFACE TEMPERATURE DISTRIBUTIONS

Several years ago Lindzen and Nigam (1987) sought to investigate whether pressure gradients directly induced by horizontal temperature gradients at the surface which are turbulently communicated to the overlying air could account for any significant portion of the low level winds. The importance of this question lies in the ability of low level winds to converge moisture, and thus establish patterns of cumulonimbus convection and rainfall, as well as the ability of such winds to drive ocean circulations. While the latent heat associated with the resulting rainfall will undoubtedly drive large scale circulations aloft, there is no compelling reason to suppose that the resulting motion will dominate the low level flow. Such a situation was suggested by the work of Schneider and Lindzen (1977) and Schneider (1977) on Hadley circulations. They found an approximate two cell structure in the vertical with the lower cell forced by surface temperature gradients and the upper cell forced by latent heating.

LN found that over much of the Pacific, the pressure gradients induced by surface temperature distributions were close to the observed pressure gradients (*viz* Figure 3). Moreover, the eddy component of the resulting flow was in fair agreement with the eddy component of the observed low level flow (*viz* Figure 4). Agreement with the observed zonally averaged component of the flow was worse, though calculated and observed components of the zonally averaged convergence were in pretty good agreement (*viz* Figure 5). More recently, DaSilva (1990) has shown that pressure gradients and low level winds induced by surface temperature gradients are also in good agreement with observations off the coast of Brazil.

In connection with our prospects for modeling the tropics, there are several remarks to be made about the above results. First, even close replication of observed

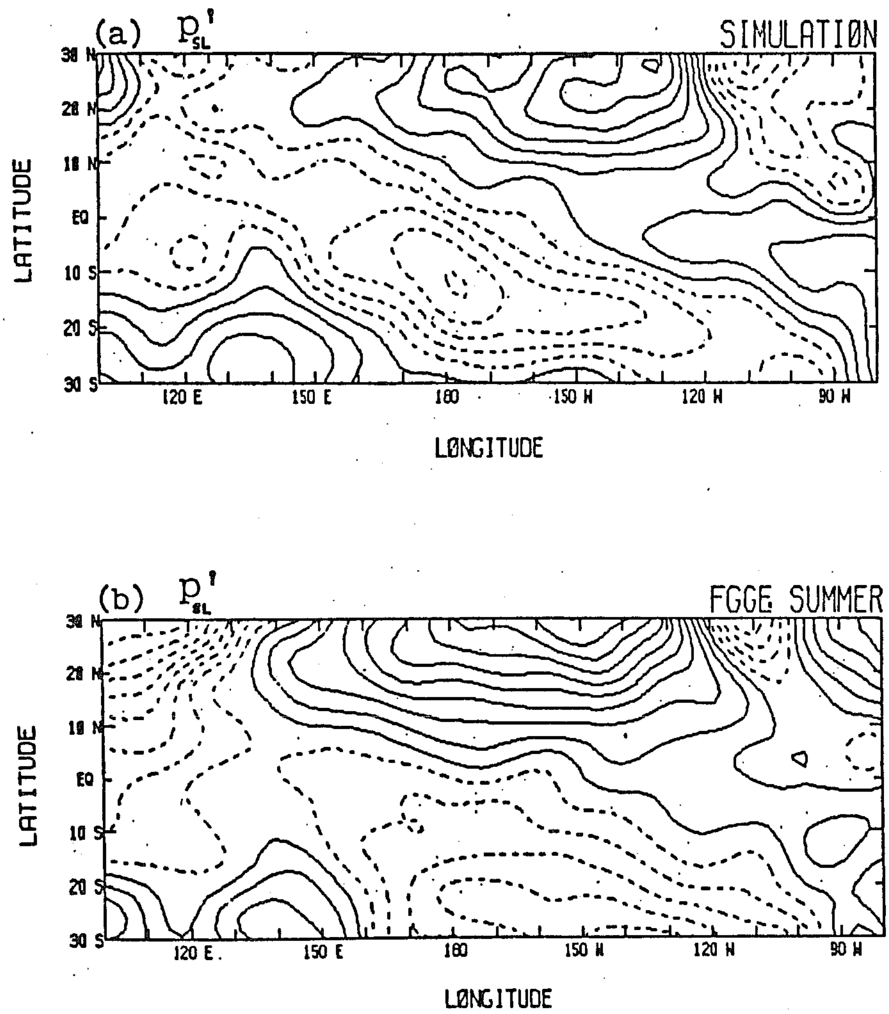

Figure 3. The calculated (a) and observed (b) eddy sea-level pressure over the Pacific basin. The observed field is obtained from ECMWF analyses for the FGGE summer. The contour interval is 1 mb. Dashed lines show negative contours; the first solid contour corresponds to zero. From LN.

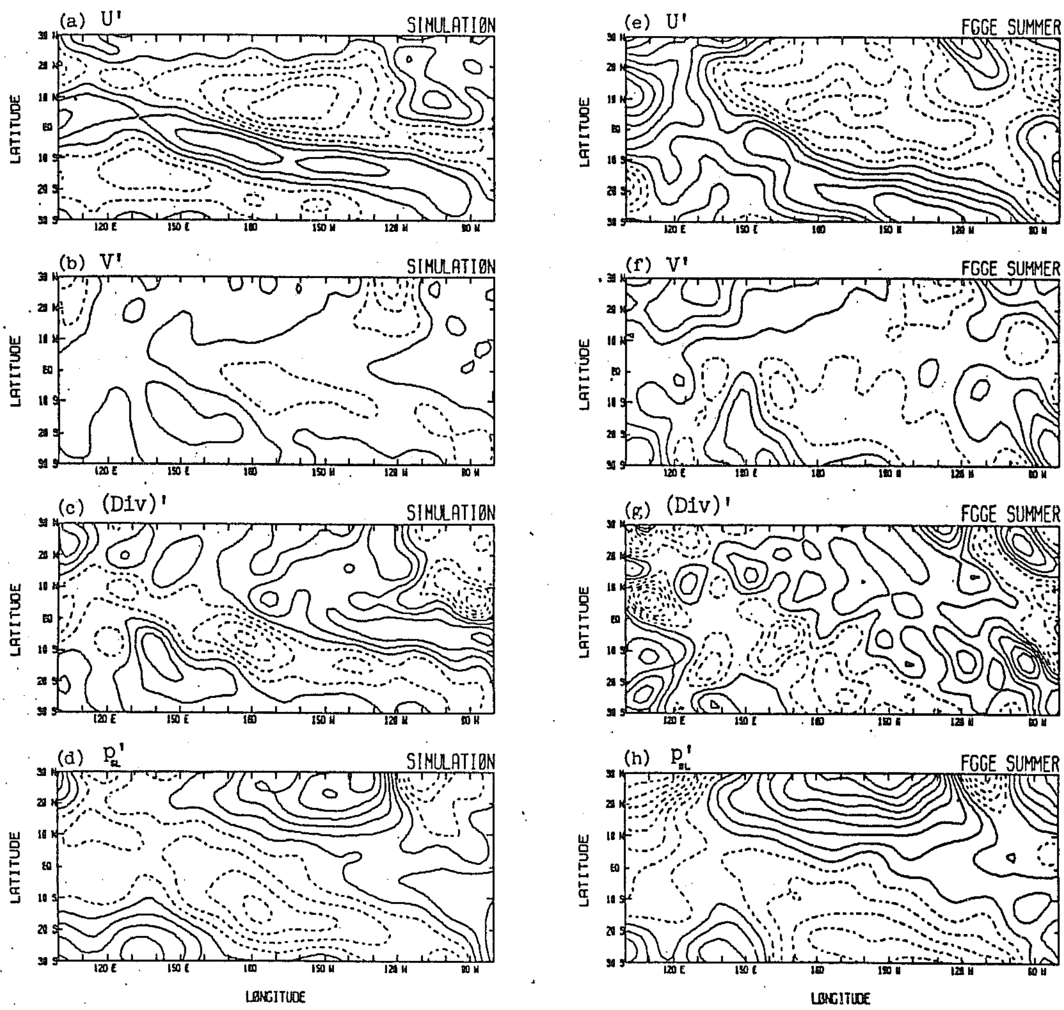

Figure 4. The low-level summertime flow over the Pacific obtained from the linear model in which the boundary layer height is allowed to adjust to the horizontal convergence. The model solutions are shown in the left panels while the corresponding fields from the ECMWF analyses are shown the right panels. The contour intervals are as follows: U' and V' = 1 m s^{-1}, div = 4 10^{-7} s^{-1}, and p'_{SL} = 1 mb. Contouring convention is as in Figure 3. From LN.

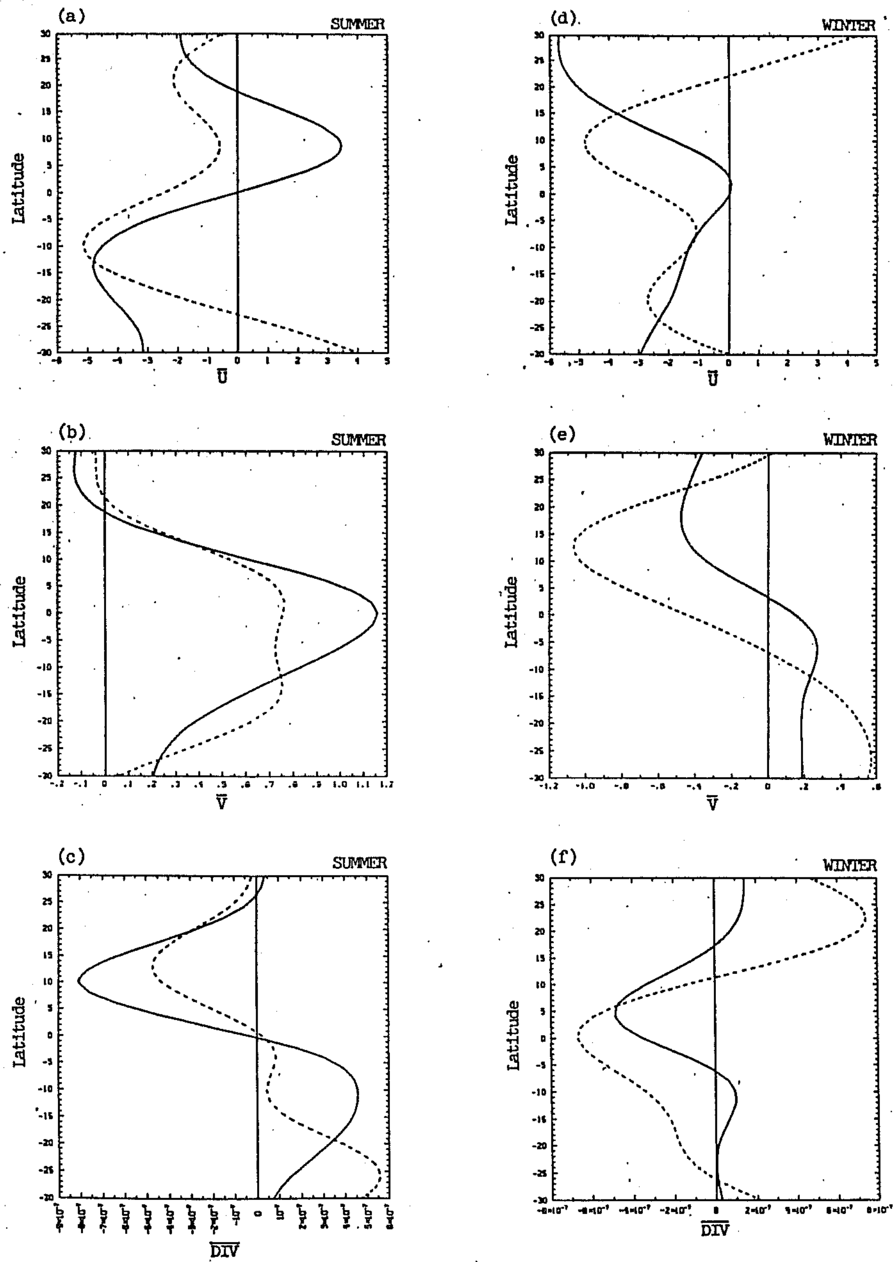

Figure 5. The low-level zonally symmetric circulation obtained using the model described in the text is shown for FGGE summer on the left and FGGE winter on the right. The dashed line in each figure is a smoothed version of the corresponding ECMWF analyzed field. From LN.

pressure distributions may not be good enough. Figure 6 shows two pressure distributions and their associated divergence fields. The two pressure fields are almost indistinguishable, but the divergence fields are radically different in the immediate neighborhood of the equator. The differences between the two cases are of considerable importance. The unrealistic case was one where we assumed a convectively mixed boundary layer extending up to a fixed 'trade inversion' at 700 mb. Convergence was assumed to be taken up instantly by cumulonimbus towers. In the more realistic case, we allowed a delay of about 30 minutes in the cumulonimbus takeup of convergence -- during which time convergence was permitted to raise the depth of the mixed layer, thus creating a 'back-pressure.' The depth change, which did not exceed 10 gpm, was enough to cause the differences seen in Figure 6. This brings us to a second point: namely, in the tropics, at least, the modeling of cumulonimbus convection and the modeling of the trade cumulus boundary layer are tightly related. The final point is really a variant on the second point: namely, the existence of effective mixing in the bottom 2-3 km of the atmosphere in the tropics is essential to the plausible modeling of tropical rainfall distributions, and the proper modeling of rainfall distribution is essential to the calculation of the circulation above the mixed moist layer.

4. ORGANIZATION OF RAINFALL BY EASTERLY WAVES

The organization by easterly waves of tropical rainfall suggested to some (Lindzen, 1974, among many others) that such waves might, in fact, be driven by the latent heat released by the rain. This process came to be known as Wave-CISK. As a mechanism for accounting for easterly waves, it suffered from a number of severe drawbacks. For one thing, without special *ad hoc* corrections, the mechanism provided no scale selection. The earliest studies found maximum growth for zero scale. Later studies (Stevens and Lindzen, 1979) found neutrality for all but the shortest scales. Observed scales were not regarded as all that mysterious. As Burpee (1972) pointed out, easterly waves generally originated in regions of barotropic instability, and the observed scales corresponded plausibly to the scales of the most rapidly growing barotropic instabilities. However, the question remained as to exactly how these barotropic instabilities actually organized the rainfall. Presumably, the instabilities must contribute sufficient convergence to the moist boundary layer to organize the rainfall. If Stevens and Lindzen (1979) are correct, then

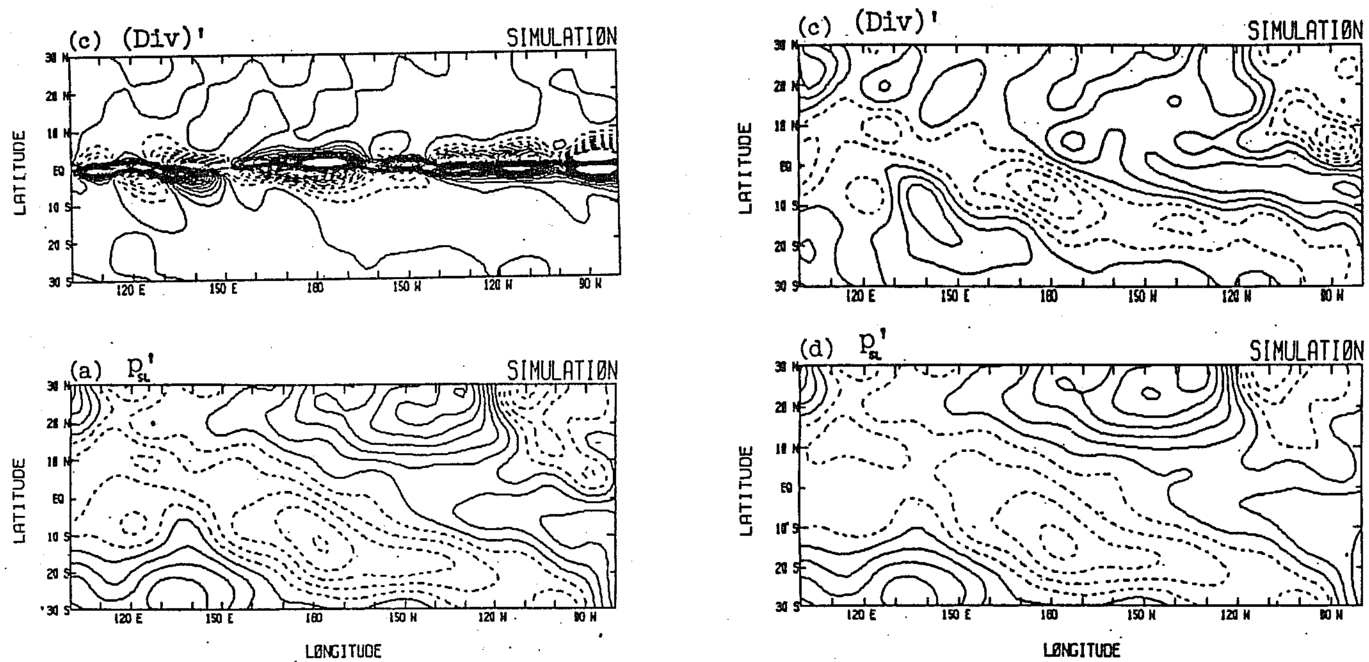

Figure 6. Pressure fields and their corresponding divergence fields. The panels on the left show results without back-pressure corrections; the panels on the right include back-pressure corrections. From LN.

the rainfall, itself, will provide enough convergence at any scale to provide neutrality. A small boost from barotropic instability may be sufficient to cause its scale to dominate.

This small boost cannot be infinitesimal since infinitesimal boosts will be associated with infinite growth times. Observations also argue for a finite threshold. For example, during all three phases of GATE the rainfall was about the same (Woodley et al, 1980). Also, during each phase, there was comparable easterly wave activity. However, only during phase 3, did the easterly waves organize the rainfall (Reeves et al, 1979). Presumably, the easterly waves during each phase contributed something to the low level convergence, but only during phase 3 was the contribution sufficient to affect the pattern of rainfall. In order to get a better idea of what distinguished phase 3 from the preceding phases, Miller and I examined the ability of waves above the moist layer to induce ascent at the top of the moist layer.

Our approach was rather simple minded. We assumed the presence of a wave of specified scale and magnitude (corresponding to observed values), and, using linear quasi-geostrophic theory, calculated the ability of such a wave to induce ascent some distance below. On the face of it we were merely reproducing the calculations of Charney and Drazin (1961) for somewhat different circumstances. However, we soon discovered that the nature of our answer was going to be rather different. The reason for this can be seen by considering the highly oversimplified situation where the mean flow is independent of height. In this case we have simple wave (or exponential) behavior with the following index of refraction

$$m^2=\frac{N^2}{f^2}\left(\frac{\beta}{U_0-c}-k^2-l^2\right)-\frac{1}{4H_0^2} \qquad (3)$$

where the variables have there usual meaning. Vertical propagation (i.e., $m^2 > 0$) is associated with the quantity (U_0 - c) being small and positive, or, more precisely

$$0<U_0-c<\frac{\beta}{k^2+l^2+\frac{f^2}{4H_0^2N^2}} \qquad (4)$$

This is, of course, exactly what Charney and Drazin got. If we choose f and β to be characteristic of 15°N, and set k=l=2 10^{-6} m^{-1} (corresponding to a zonal wavelength of about 30° of longitude), we find that propagation requires that U_0-c must be less than 2-3 m/s. For downward propagating solutions where

$$\Phi = \Phi_0 e^{-im(z-z_T)}, \tag{5}$$

the wave's contribution to ascent (i.e., vertical velocity) is given by

$$|w| = \frac{k}{N^2}(U_0-c)me^{(z-z_T)/2H_0}|\Phi_0|. \tag{6}$$

Now, for propagating solutions, Equation 6 has a maximum, given by

$$|w| = \frac{\beta}{2Nf}\frac{k}{\sqrt{k^2+l^2}}e^{(z-z_T)/2H_0}|\Phi_0| \tag{7}$$

which occurs for

$$U_0-c = \frac{\beta}{2(k^2+l^2)} \tag{8}$$

The trouble with the result given by Equation 7 is that if we take Φ_0 = 23.5 m^2s^{-2} (corresponding to a meridional velocity of 1.25 ms^{-1}), then the maximum value of |w| obtained from Equation 7 is 0.17 mb hr^{-1}, which turns out to be so much smaller than the observed ascent of 3-5 mb hr^{-1} so as to be incapable of organizing convection. As it turns out, however, the same constraint does not apply to trapped solutions. For these solutions,

$$\Phi = \Phi_0 e^{n(z-z_T)} \tag{9}$$

where n^2=-m^2, and

$$|w| = \frac{k}{N^2}(U_0-c)(n+\frac{1}{2H_0})e^{(n+\frac{1}{2H_0})(z-z_T)}|\Phi_0| \tag{10}$$

From Equation 10, we see that $|w|$ can be made arbitrarily large by making U_0-c large. On the other hand solutions decay rapidly with depth (especially so since f is small in the tropics). The conclusion one is drawn to is that, in order to organize convection, a barotropic instability must be such that U_0-c is large at the top of the moist layer, and the barotropically unstable layer must be close to the moist layer. Characteristic decay scales are on the order of 1 km, so that small vertical shifts in the barotropically unstable layer can have major effects of the ability of a wave to organize convection. All these conclusions are confirmed by calculations for more realistic profiles of U_0.

Some further features emerge from a consideration of specific profiles of U_0 appropriate to the region below the African jet. In Figure 7 we show 4 profiles that could not readily be distinguished in the data. In Figure 8 we show the profiles of w associated with each of the U_0 profiles in Figure 7, and in Figure 9 we show profiles of Φ. We see that small differences in U_0 (including changes in the flow curvature) can lead to factor 2 differences in w; we also see that it is possible to have sharp downward decay of w without accompanying decay of Φ. This last feature is consistent with the observation of Chen and Ogura (1982) that during phase 1 of GATE, wave vorticity (proportional to the geopotential for quasi-geostrophic perturbations) was virtually undiminished between the jet level and the surface, while surface convergence remained relatively unaffected by the wave.

The above calculations show that only a modest elevation (O(1 km)) in the jet during phase 1 of GATE could readily account for a major reduction in the ability of the wave to organize convection. Such an elevation over the GATE ship array was reported by Chen and Ogura (1982). Assuming this to be the case, it would appear that the smallest contribution of the easterly wave to the ascent at the top of the moist layer during phase 1 would be about 0.7 mb hr^{-1}; presumably, some such value constitutes the threshold beyond which the wave contribution to ascent must go in order to significantly organize rainfall. As we have already noted, this is much greater than the largest possible contribution from vertically propagating waves.

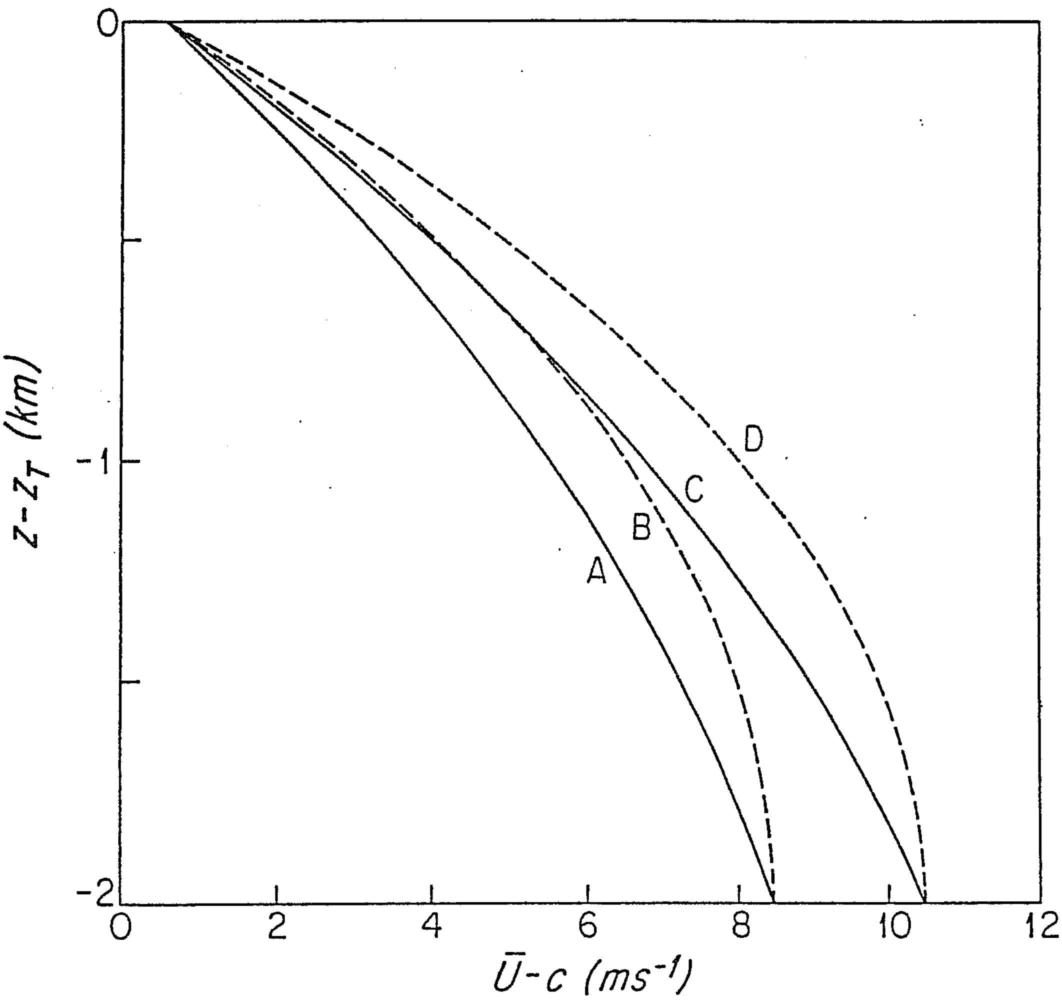

Figure 7. Four zonal wind profiles chosen to resemble the African Jet. The profiles are quadratic in z, and characterized by their mean shear, $\Delta U/\Delta z$, and by U_1, the shear at z_T. For curve A, $\Delta U/\Delta z$=-4m/s/km, and $U_1=1.5\Delta U/\Delta z$; for curve B, $\Delta U/\Delta z$=-4m/s/km, $U_1=2\Delta U/\Delta z$; for curve C, $\Delta U/\Delta z$=-5m/s/km and $U_1=1.5\Delta U/\Delta z$; and for curve D, $\Delta U/\Delta z$=-5m/s/km, and $U_1=2\Delta U/\Delta z$. Δz=2 km.

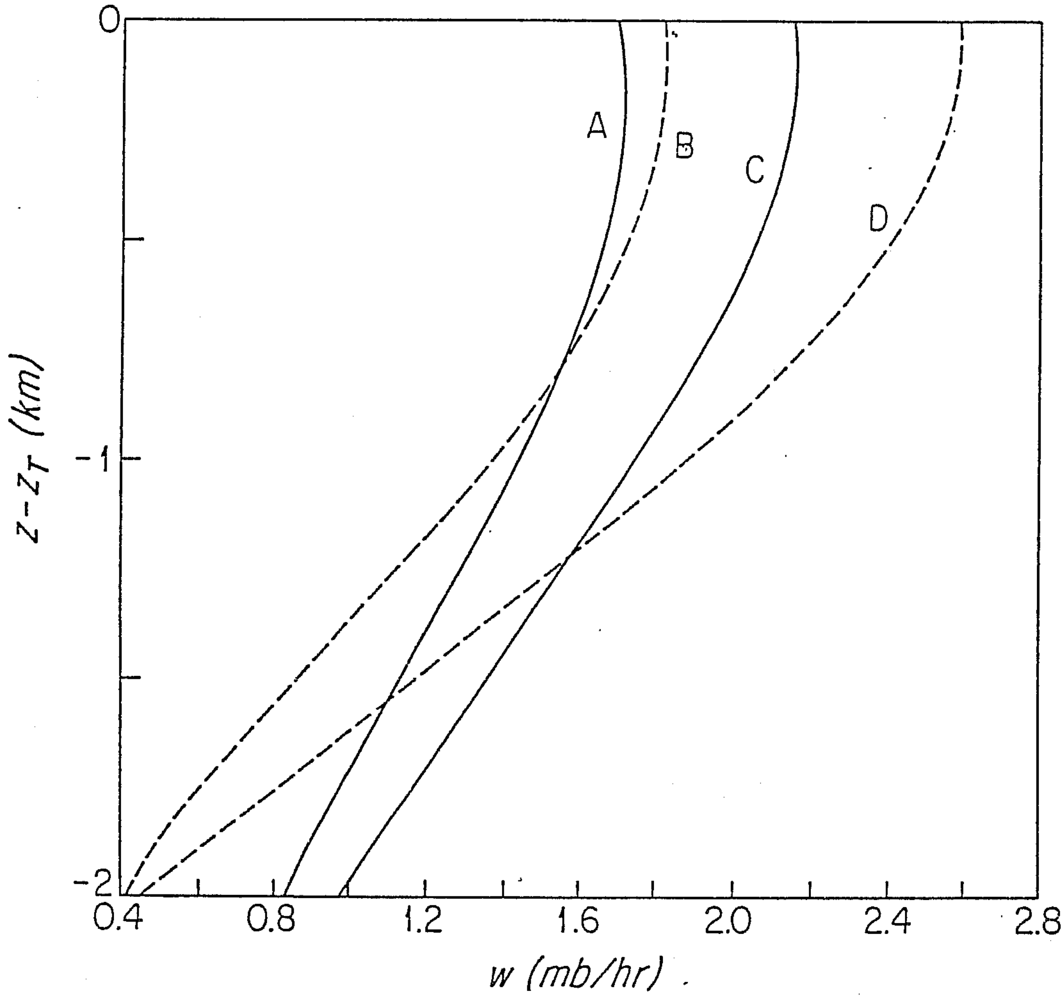

Figure 8. The ascent in mb hr^{-1} corresponding to the four profiles in Figure 7.

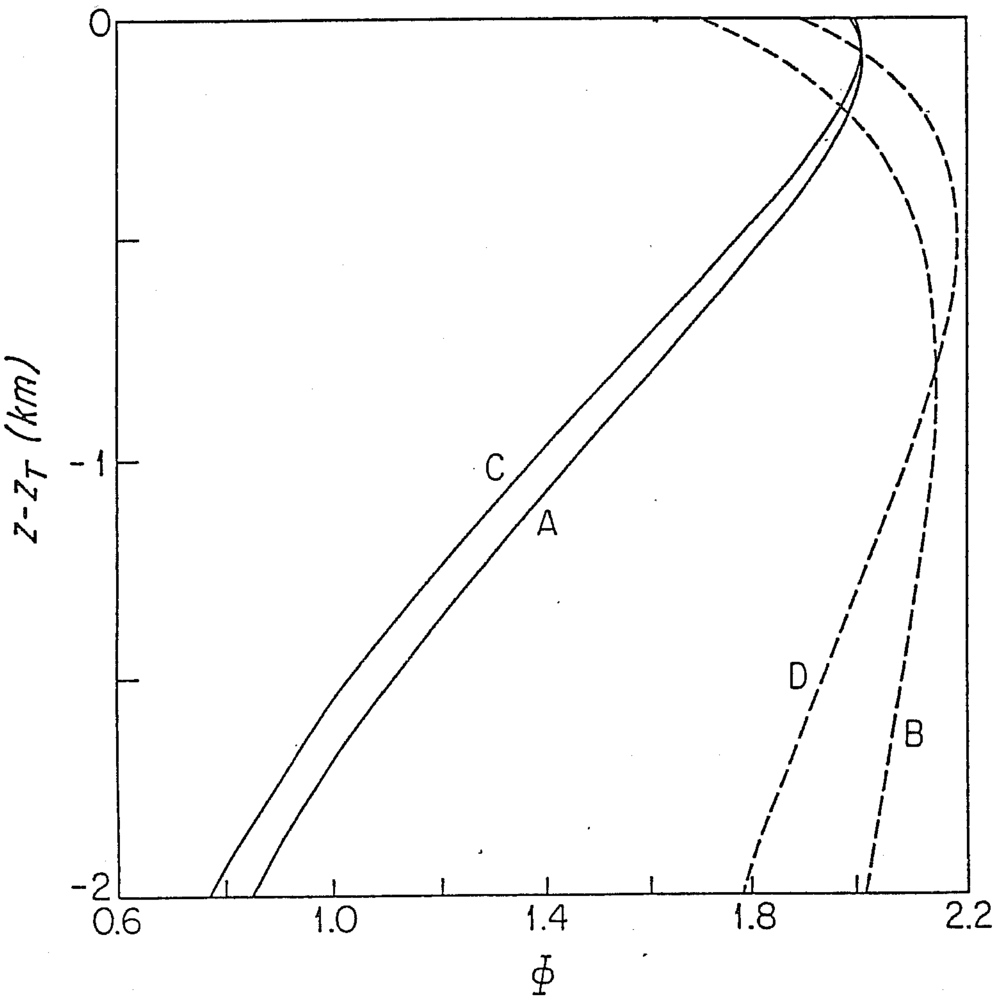

Figure 9. The geopotential (normalized by the amplitude of the downgoing wave at z_T) corresponding to the four profiles in Figure 7.

5. PROSPECTS FOR TROPICAL MODELING

The above sections have highlighted three specific aspects of tropical meteorology which place rather important demands on models -- if the models are to be successful. Considerations of consistent resolution suggest that vertical resolution much greater than present in existing models may be needed -- at least for long term integrations. Results in section 4 suggest that vertical resolution in both the model and the data need to be appreciably better than 1 km in order to adequately deal with easterly waves. These results suggest a sensitivity to curvature in the basic flow which is unlikely to revealed by 'significant levels.' Section 3 shows a profound sensitivity to extremely small pressure gradients in the neighborhood of the equator. However, proper treatment of the physics seems likely to markedly reduce this sensitivity. 'Proper physics' seems to include some representation of the fact that the lowest 2-3 km are significantly mixed by trade cumuli, and some means for accounting for the time dependence of cumulonimbus development.

Acknowledgements

This paper was prepared with support from the National Science Foundation under Grant 8520354-ATM, and from the National Aeronautics and Space Administration under Grant NAGW 525.

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