Diurnally Forced Tropical Tropospheric Circulation over South America

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(Manuscript received 31 August 1985, in final form 31 October 1986)

ABSTRACT

The impact of diurnally varying thermal forcing on the tropical circulation is studied with a linearized shallow water model on an equatorial beta plane. The mass and flow fields are interpreted as an internal mode with equivalent depth of the order of 250 m. This is the preferred vertical mode excited by typical cumulus heating profiles in the tropics. The diurnal forcing with the horizontal scale of the Amazon source partitions approximately 40% of the input energy to slow modes. The fast modes are trapped by the critical latitudes contributing significantly to oscillations in the divergence field, which emanate out of the source region. The divergence field at the source is approximately in phase with the forcing; the vorticity field lags the forcing by approximately 10 hours. A comparison of the model results with the 500 mb level vertical motion, derived from the National Meteorological Center analysis, shows a similar pattern of propagating oscillations. It is suggested that the diurnally forced convection over the tropical sector of South America controls the timing of heavy precipitation over the adjoining regions.

1. Introduction

The convective activity over tropical South America is primarily concentrated over the Amazon Basin and Central Brazil during the summer season (Ratisbonna, 1976) and is clearly modulated by the diurnal heating cycle (Minnis and Harrison, 1984). The convective forcing over the region is also affected by other time scales which seem to be correlated with midlatitude systems which penetrate into the tropical area as discussed by Kousky (1979), Kousky and Ferreira (1981) and Virji and Kousky (1983). From a modeling point of view, the response of the upper tropospheric circulation over South America to transient convection has been recently studied by Silva Dias et al. (1983). The latter discussed the simulation of the upper tropospheric response to a Gaussian-shaped transient mass source using an equatorial β -plane shallow-water equation model assuming an equivalent depth (D) on the order of 250 m. The particular equivalent depth assumed by Silva Dias et al. (1983) has also been used by Gill (1980), Lim and Chang (1981) and Lau and

Lim (1982) and it is consistent with an internal mode with an inflow–outflow vertical profile with zero crossing in the middle troposphere. The model results are representative of the upper tropospheric anticyclonic circulation over Bolivia, which is a predominant feature of the summer circulation over tropical South America. The transient forcing was assumed to have a time scale on the order of a few hours to a few days and the energetics problem was discussed in terms of the amount of energy which was generated in Rossby, Kelvin and gravity waves. Evidently, short period forcing resulted in a larger partition of energy towards Kelvin and gravity modes which might have a significant role as a short term teleconnection mechanism in the tropical region. Forcing functions with longer time scales tend to favor Rossby and mixed Rossby–gravity waves, which are responsible for the quasi-balanced anticyclone and the well developed trough off the northeast coast of Brazil, studied by Kousky and Gan (1981).

The observational evidence of the diurnal modulation of convection over tropical South America (Kousky, 1980; Minnis and Harrison, 1984), raises the possibility of studying the effects of a diurnally varying mass source on the simple equatorial β -plane model of Silva Dias et al. (1983). Further observational evidence of the diurnal cycle, based on the difference between 0000 and 1200 UTC vertical motion fields at 500 mb over the tropical South American sector, is discussed in section 2.

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** The National Center for Atmospheric Research is sponsored by the National Science Foundation.

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The main goal of the present study is to analyze possible influences that a localized diurnal convective forcing, such as the forcing over the Amazon region, has on the large scale. The importance of diurnal convective forcing was noted in observational studies (Foltz and Gray, 1979) and also in the model results of Silva Dias et al. (1983). It will be shown that the diurnal source excites fast modes that produce trains of convergence-divergence patterns that propagate primarily along the equator. The associated divergence has some resemblance to the vertical motion field calculated using the tropical National Meteorological Center (NMC) final analysis. The choice of a single equivalent depth analysis of the equatorial β -plane model of Silva Dias et al. (1983) can be questioned because of the rigid top assumption, which is inherent to the type of model that is being used (Lindzen et al., 1968; Geisler and Stevens, 1982). However, most of the circulation characteristics, generated by a diurnal mass source, are already present after two or three cycles. Therefore, we expect little effect from a rigid-top boundary on such a time scale. Alternatively, one can interpret the results as being representative of most numerical models which have some sort of rigid upper boundary condition. More recently, Silva Dias and Bonatti (1985) decomposed FGGE level III-b data over tropical South America into vertical modes of a linearized σ -level primitive equation model used by Kasahara and Puri (1981). Silva Dias and Bonatti show the dominance of an internal mode, with equivalent depth of the order of 220 m over the convectively active tropical region of South America. Similar results were obtained in other parts of the tropics by Silva Dias and Paegle (1984). The same model has been used by Puri (1983) to analyze the effect of the convective parameterization on the generation of the Hadley cell. The same vertical mode with equivalent depth of the order of 220 m was also noted to be important in Puri's model results. This sharp spectral peak in the vertical mode projection was also found by Fulton and Schubert (1985). Thus, the choice of just one representative vertical mode is justified.

A short description of the linearized shallow-water model developed in Silva Dias et al. (1983) is given in section 3 along with the formal solution for the diurnal mass source. The model results for an oscillatory gaussian shaped mass source located at approximately 15°S with an e -folding half-width of 1500 km are shown in section 4.

2. Observational evidence

Satellite imagery provides us with perhaps one of the best means to analyze the diurnal convective cycle over South America. As an example, we have selected the images shown in Fig. 1. For the period shown (01/28/81 to 01/30/81), the images on the left hand side of the figure were taken near 1200 UTC (0800 LT in

the central Amazon basin) at a time when convective activity is near a minimum. Those on the right hand side were taken near 00 UTC (time in two-digit hours), which corresponds to the time when convective activity is near its peak (20 LT). The diurnal variation in the quantity of cold cloud tops (white areas in the images), representative of deep convection, is readily apparent.

Recently, a method has been developed by which precipitation amounts can be estimated using the geostationary infrared data (Arkin, 1983). An example of this is shown in Fig. 2a for the southern summer, during 1983 (Phillip Arkin, personal communication, 1983). The amplitude and the time of maximum contribution (local time) of the first harmonic of the diurnal cycle in precipitation, based on the method of Arkin, are shown in Figs. 2b and 2c, respectively. Some interesting features in Fig. 2c are the late afternoon maximum in precipitation over most of Brazil, the early morning maxima over central Argentina, over the west coast of South America near Ecuador and Colombia and over the equatorial Atlantic near 30°W.

Since there is considerable interannual and even month-to-month variation in the spatial distribution of deep convection, during the summer over South America, mean wind fields taken over several years or even over a summer season would be expected to show a diurnal cycle weaker than that for shorter periods. Since we are interested in relatively short term responses to thermal forcing, we have sought periods during which the convective regime over South America was highly persistent. In choosing these periods, a subjective analysis of geostationary infrared satellite imagery was made. The diurnal variation in the wind field, for one of these periods (January 1981) was obtained from the NMC tropical analyses.

The average vertical motion in pressure coordinate (ω) at 500 mb at 00 and 12 UTC is shown in Fig. 3a and 3b, respectively. The divergence is linearly mass balanced in order to assure zero vertical motion at 1000 and 200 mb. The mean field, defined as the arithmetic mean of the 00 and 12 UTC fields, is shown in Fig. 3c. The perturbation vertical motion field, defined as the difference between the 00 UTC field (Fig. 3a) and the mean field (Fig. 3c), is shown in Fig. 3d. The statistical significance of the perturbation field shown in Fig. 3d was verified with the standard t -test. The null hypothesis is that the samples obtained from the 00 and 12 UTC data came from the same population. The " t " parameter was computed at each grid point and the shaded area indicates the regions which can not be rejected at the 95% confidence level.

The perturbation vertical motion at 00 UTC (Fig. 3d) shows a negative deviation (upward motion) over central South America with an extension into the Atlantic. Relative sinking occurs along the equator in Fig. 3d with three main cells located at approximately 75°, 40° and 15°W. Just north of the equatorial band of sinking motion perturbation, there is a band of upward

FIG. 1. Satellite infrared imagery of the South American and Atlantic region indicating the diurnal variation for the indicated dates (a), (b), (c) and (d).

vertical motion perturbation with maxima at the same longitude as the downward cells in the equatorial region. At subtropical latitudes in the Southern Hemisphere there is also an east-west belt of perturbation sinking motion which extends from the coast of Chile to the Atlantic coast of Southern Brazil. The perturbation vertical motion field for 12 UTC has the same pattern as shown in Fig. 3d except that the signs are all reversed. The perturbation divergence pattern at 200 mb (not shown) is in very close agreement with Fig. 3d.

The mean relative vorticity field at 200 mb for January 1981 is shown in Fig. 4a as well as the associated wind vectors. The dashed lines in Fig. 4a represent

cyclonic relative vorticity in the Southern Hemisphere (negative values). The corresponding perturbation field for the 12 UTC observation time is shown in Fig. 4b (arithmetic mean of the 00 and 12 UTC values minus the 12 UTC observation). The average field for January 1981 captures most of the common features of the upper circulation over tropical South America such as the upper Bolivian High, the trough off the northeast coast of Brazil and the strong subtropical jet to the south of the Bolivian High. The perturbation field (Fig. 4b) is much noisier than the average field (Fig. 4a) but reveals some interesting characteristics: (i) a general tendency for a stronger anticyclonic flow over tropical South America at 12 UTC (08 LT), (ii) large magnitude

FIG. 2. (a) The GOES precipitation index in cm from December 1982 to February 1983. (b) Amplitude (cm) and (c) phase of the first harmonic of the diurnal cycle (local time) for the Southern Hemisphere summer, 1983. (Phillip Arkin, personal communication)

wind perturbation vectors over the western Amazon region (southeasterly anomaly), and (iii) equatorward flow along the northern coast of Brazil with opposite direction in each hemisphere. However, the statistical significance is much lower than in the case of the vertical motion field shown in Fig. 3. Possible reasons for the lower statistical significance will be discussed in section 4.

The high statistical significance of the main features of the vertical motion perturbation field in the tropical

region is of interest (Fig. 3d). In fact, most of the main features are significant at the 99% confidence level. The high significance is a result of the regularity of the diurnal cycle during the analyzed period. A similar analysis, on monthly averages on the regular tropical NMC grid, usually does not reveal similar significance although the general vertical motion perturbation patterns are present. However, a composite technique with respect to the center of the Bolivian High does produce a significant diurnal signal (Santos and Silva Dias, 1984).

FIG. 3. Vertical p -velocity (10 Pa s^{-1}) at 500 mb for January 1981: (a) for 00 UTC, (b) for 12 UTC, (c) mean (00 and 12 UTC) and (d) perturbation (00 UTC minus 12 UTC). Dashed lines represent negative values. The hatched area in (d) is statistically significant at the 95% level.

3. Model description

Silva Dias et al. (1983) consider a compressible hydrostatic fluid on an equatorial β -plane model, linearized about a basic state at rest and which is forced by known sources/sinks of heat (Q). The method of solution is based on the application of a finite Sturm–Liouville transform in order to decouple the vertical

dependence. This can be accomplished provided the kernel of the transform satisfies a Sturm–Liouville problem usually known as the vertical structure equation, with D_n ($n = 0, 1, \dots$) as eigenvalues, also known as the equivalent depths. Figure 5 shows the vertical eigenfunctions and eigenvalues for a basic state representative of the tropical atmosphere over South America (taken from Silva Dias and Bonatti, 1985).

FIG. 4. Relative vorticity and wind at 200 mb for January 1981; (a) mean (00 and 12 UTC) and (b) perturbation (00 UTC minus 12 UTC). Dashed lines represent negative relative vorticity and the wind speed is proportional to the length of the arrows as indicated.

The regular σ -coordinate is used in the vertical, as in Kasahara and Puri (1981). The vertical boundary conditions are given by zero vertical velocity ($\sigma = 0$) at the top (0 mb) and bottom of the model atmosphere. The application of the vertical transform to the original set of equations yields the governing equations for each

vertical mode. The resulting equations are the shallow water equations with pure gravity wave speed $c_n = (gD_n)^{1/2}$ where g is gravity,

$$\frac{\partial u_n}{\partial t} - \beta y v_n + \frac{\partial P_n}{\partial x} = 0 \quad (3.1)$$

$$\frac{\partial v_n}{\partial t} + \beta y u_n + \frac{\partial P_n}{\partial y} = 0 \quad (3.2)$$

$$\frac{\partial P_n}{\partial t} + c_n^2 \left(\frac{\partial u_n}{\partial x} + \frac{\partial v_n}{\partial y} \right) = \frac{\partial \tilde{P}_n}{\partial t} \quad (3.3)$$

The term $\partial \tilde{P}_n / \partial t$ is related to the vertical transform of the vertical derivative of the heating function, (u_n, v_n) are the zonal and meridional projections of the wind field (Hack and Schubert, 1981). P_n is the n th projection of the state variable defined by $P = \phi + \sigma \bar{\alpha} P_s$, where ϕ is the geopotential, $\bar{\alpha}$ is the specific volume of the basic state atmosphere and P_s is the surface pressure.

The heating produced by deep convective activity in the tropics (e.g., Yanai et al., 1973) is basically projected onto a single internal mode with equivalent depth on the order of 200 m, which corresponds to the fourth mode of Fig. 5 (Gill, 1980; Lau and Lim, 1982; Silva Dias et al., 1983; Fulton and Schubert, 1985). Vertical decomposition of FGGE Level III-b data has also shown that the fourth mode dominates the energetics of the tropical active regions (Silva Dias and Bonatti, 1985). Thus, we will assume that the fourth mode is representative of the solution of (3.1)–(3.3) under the typical tropical heating source/sink function and ignore the contribution of the remaining vertical modes.

Under the above simplification, it is convenient to convert (3.1)–(3.3) to the nondimensional form with $\{L\} = \{c/\beta\}^{1/2}$ and $\{T\} = 1/(c\beta)^{1/2}$ as the appropriate length and time scales, respectively (Matsuno, 1966). Using the same symbols for the nondimensional system of equations, we have, in vector form,

$$\frac{\partial \zeta}{\partial t} + \mathbf{L} \zeta = \mathbf{F} \quad (3.4)$$

where

$$\zeta = \begin{bmatrix} u \\ v \\ P \end{bmatrix}, \quad \mathbf{L} = \begin{bmatrix} 0 & -y & \frac{\partial}{\partial x} \\ y & 0 & \frac{\partial}{\partial y} \\ \frac{\partial}{\partial x} & \frac{\partial}{\partial y} & 0 \end{bmatrix}, \quad \mathbf{F} = \begin{bmatrix} 0 \\ 0 \\ \frac{\partial \tilde{P}}{\partial t} \end{bmatrix} \quad (3.5)$$

The horizontal boundary conditions are $\zeta \rightarrow 0$ when $y \rightarrow \infty$ and zonal periodicity associated with the equatorial circumference of the earth. The method of solution of (3.4) is based on the expansion of (3.5) in a convergent series which transforms (3.4) into a system

FIG. 5. Vertical structure of eigenfunctions corresponding to the equivalent heights D , $n = 1, \dots, 9$ for the 9-level finite difference version of the vertical structure equation in Kasahara and Puri (1981).

of ordinary differential equations for the expansion coefficients. The basis functions are chosen as the eigenfunctions of the linear operator L which form a complete and orthogonal set on the equatorial β -plane (Cane and Sarachik, 1977). The eigenfunctions and eigenfrequencies ($\omega_{k,m,r}$ and $\zeta_{k,m,r}$, respectively) are dependent on the indices (k, m, r), where k is the zonal wavenumber, m is the meridional index and $r = 0, 1, 2$. When $r = 0$ we have the low frequency Rossby waves; the higher frequency gravity waves propagate toward the west and east, corresponding to $r = 1$ and 2, respectively. The meridional structure is related to the index m . Thus, the solution for (3.4) is given by the series

$$\zeta(x, y, t) = \sum_k \sum_m \sum_r c_{k,m,r}(t) \zeta_{k,m,r}(y) e^{ikx} \quad (3.6)$$

and the expansion coefficients are governed by

$$\frac{d}{dt} c_{k,m,r} - i\omega_{k,m,r} c_{k,m,r} = f_{k,m,r} \quad (3.7)$$

where $f_{k,m,r}$ is the expansion coefficient of the forcing term $\mathbf{F}$. The time dependent part of the forcing function is a half-sine during half a day (in nondimensional units) and zero otherwise, with maximum amplitude equal to one nondimensional unit, as shown in Fig. 6. The space dependent part is given by a two-dimensional Gaussian surface, with e -folding radius a and centered at a latitude y , as in Silva Dias et al. (1983). After Fourier transforming the time dependent part of the forcing, the expansion coefficients become

$$c(t) = \frac{f}{\pi} \sum_{\substack{j=2 \\ (\text{even})}}^{\infty} \frac{1}{(1-j^2)} \left[\frac{e^{i\omega t} - e^{i\pi j t/\tau}}{i\left(\omega - j\frac{\pi}{\tau}\right)} + \frac{e^{i\omega t} - e^{-i\pi j t/\tau}}{i\left(\omega + j\frac{\pi}{\tau}\right)} \right], \quad (3.8)$$

provided the initial condition is $\zeta(x, y, 0) = 0$. In (3.8) the subscripts (k, m, r) have been deleted and τ is the

duration of a day in nondimensional units. The definition of f in (3.8) can be found in Eq. (3.11) of Silva Dias et al. (1983).

4. Model results

Previous model work on the role of transient heat sources in the tropical region suggests that fast forcing excites a large amount of energy in the gravity modes (Paegle, 1978; Silva Dias et al., 1983). This feature seems to be supported from the observational point of view (section 2), since the diurnal signal in the vorticity field is marginal, compared to the divergent component during a period of intense convective activity over tropical South America. In this section, we focus on the partition of energy between fast and slow modes for the diurnal forcing. This problem will be studied with the linear primitive equation model, based on one single representative vertical mode, described in the previous section.

In order to simulate the diurnal convective forcing over tropical South America, we positioned the Gaussian-shaped mass source (section 3) at the nondimensional latitude $y = -0.75$ and with an e -folding width $a = 0.54$. Thus, the mass source is located at approximately 15°S with an e -folding radius on the order of

FIG. 6. Amplitude of the forcing function as a function of time. Duration of a day is given in nondimensional units.

800 km for the representative internal mode with $c = 50 \text{ m s}^{-1}$ ($\{L\} = 1500 \text{ km}$ and $\{T\} = 0.35 \text{ day}$). A gross comparison of these parameters, with the observed structure of the convective region over tropical South America (section 2), shows reasonable agreement. The heating function (Fig. 6) is assumed to start acting on the fluid at the initial time. After the third cycle, the wind and geopotential fields differ from the previous fields basically in magnitude, but not in shape. With the inclusion of damping, it is possible to obtain a stationary solution which repeats itself exactly at each cycle. This type of simulation was also performed with linear friction and cooling. Although some of the results are consistent with the frictionless model, the representativity of this type of solution is questionable because of multiple reflections that might be induced by the upper boundary condition.

The time variation of the total energy and the ratio between energy in rotational and gravitational modes (R) is shown in Fig. 7. The time behavior of the forcing function is also indicated in Fig. 7. As a threshold between rotational (slow) and gravitational modes (fast), we have used the convention adopted by Tribbia (1979), i.e., if $|\omega_{k,m,r}| < 2^{-1/2}$ we have the slow modes and $|\omega_{k,m,r}| > 2^{-1/2}$ defines the fast modes. Without friction, the total energy increases stepwise whenever mass is added to the system (or equivalently, when heating is applied), remaining constant between the periodic bursts. The ratio R tends toward a limit as the number of cycles increases. At the end of the fourth cycle, ap-

proximately 40 percent of the total energy is in rotational modes.

Figure 8 shows the wind and geopotential fields for the fourth cycle (Fig. 6) at cycle times $t = 0$ (Fig. 8a, just prior to the initializing forcing), $t = \tau/4$ (Fig. 8b, the maximum of the heating cycle), $t = \tau/2$ (Fig. 8c, the forcing has just subsided), $t = 3\tau/4$ (Fig. 8d, midway between one heating cycle and the beginning of the next) and $t = \tau$ (Fig. 8e, the beginning of the next heating cycle). The associated divergence and vorticity are shown in Figs. 9 and 10, respectively. The nondimensional scales are shown on the bottom and left ordinates; the dimensional scales for the fourth mode of Fig. 5 are on the right and top of Figs. 8, 9 and 10. The interpretation of the model results in terms of the internal mode, with $c = 50 \text{ ms}$ and the vertical structure shown in Fig. 5, implies that the dynamical fields shown in Figs. 8–10 can be interpreted as being representative of the upper troposphere (approximately 200 mb) or lower troposphere, provided the signs are reversed. However, the circulation in the lower troposphere has smaller amplitude, according to the vertical structure of the $n = 4$ vertical mode of Fig. 5. The center of the forcing is indicated by the dot symbol at the center of Figs. 8–10.

One may interpret the model derived divergence fields in Fig. 9, in a qualitative sense, as either the divergence field in the upper troposphere (200 mb) or as the vertical motion of the middle troposphere. Thus, comparison between the model derived divergence and the observed diurnal variation of the vertical motion is possible (comparison between Fig. 9 and Fig. 3). The 00 UTC perturbation field (Fig. 3d for vertical motion perturbation and Fig. 4b for vorticity perturbation) is representative of the situation a few hours after the maximum convective activity in the Amazon region, as shown in the satellite derived phase of the diurnal precipitation cycle (Fig. 2c). The 12 UTC perturbation fields corresponds to the period of minimum convective activity over the area. The time behavior of mass source (Fig. 6) indicates that $t = \tau/4$ (Figs. 8b, 9b and 10b) and $3\tau/4$ (Fig. 8d, 9d and 10d) closely correspond to 00 and 12 UTC, respectively.

At the beginning of the convective cycle, we have an anticyclonic circulation with maximum southerly wind on the eastern side of the cell (Fig. 8a). The flow turns to easterly just south of the equator. The main anticyclonic cell is in approximate geostrophic balance; the remaining pressure cells have no clear relationship between the mass and wind fields and change quite rapidly in time, this being a manifestation of gravity waves.

It is interesting to note that the horizontal wind distribution changes as the forcing intensifies: the flow just south of the equator becomes southeasterly (Fig. 8b) and the southerly winds to the east of the forcing disappear by $t = \tau/2$ (Fig. 8c). Similar features can be noted in the observational data (Fig. 4). From this time

FIG. 7. Total energy (solid line) with scale on the left in nondimensional units and ratio of rotational energy to total energy (dashed line) as a function of time (scale on the right in % of total energy). The results presented in Figs. 8–10 are obtained at the fourth cycle as indicated by the amplitude of the forcing (dotted line).

FIG. 8. Wind and geopotential fields at time $t = 0$ (a), $\tau/4$ (b), $\tau/2$ (c), $3\tau/4$ (d) and τ (e) as indicated in Fig. 6. The symbol ● represents the center of the forcing function which has an e -folding width of approximately 0.5 non dimensional units. Scales at bottom and on the left are nondimensional and the right and top scales are dimensionalized based on the equivalent depth of the fourth internal mode (Fig. 5). Dashed lines indicate negative values.

on, a low pressure center develops at $x = 10.3$ and $y = -0.7$ (indicated by the letter A in Fig. 8c), and the southerly flow on the eastern flank of the main anticyclonic cell intensifies. This low pressure center intensifies between $t = \tau/2$ and $t = 3\tau/4$, because more

mass is taken out of the region by the strong easterlies just south of the equator than is added by the southerly flow. It is this low pressure system that evolves to the two centers just north and southeast of the main high A_1 and A_2 , respectively) at $t = 0$ and $t = \tau$ (Fig. 8a and

FIG. 9. As in Fig. 8 but for the divergence field.

8e, respectively). As time increases, the low pressure in the Northern Hemisphere splits in two parts with one ($A_{1,1}$) moving westward and maintaining its strength and with the eastward bound ($A_{1,2}$) moving further north (Fig. 8b). The centers indicated by $A_{1,1}$

and $A_{1,2}$ can still be tracked as $A'_{1,1}$ and $A'_{1,2}$ in Fig. 8c, Fig. 8d and Fig. 8e. The Southern Hemisphere center A_2 can also be tracked as A'_2 in its eastward trajectory towards the equatorial region (Fig. 8e).

The fast moving high pressure centers can also be

FIG. 10. As in Fig. 8 but for the relative vorticity field.

tracked as they emanate from the source region along the tropical channel. The main center is indicated by the letter B in Fig. 8c, developing in response to the mass accumulation just north of the source due to the cross equatorial forced flow during the active period

of the source. In a manner analogous to the low pressure center in Fig. 8a, the B cell splits into two parts (westward and eastward bound, respectively). The centers can be tracked up to the time when they leave the displaying domain. In fact, all pressure centers asso-

ciated with the fast moving gravity waves have similar behavior, which can be explained in terms of the ray-path theory applied to the gravity waves. These modes are confined to the equatorial region, since gravity energy is trapped by the critical latitudes where the frequency of the mode equals the local Coriolis parameter (Gill, 1982).

The quasi-stationary features of the solution (associated with the rotational modes) also show a strong diurnal variation as seen in Fig. 8a to 8e; the center of the upper anticyclonic circulation moves closer to the mass source region as the forcing reaches its maximum value at $t = \tau/4$ (Fig. 8b), and then slides to the southwest. On the southeast side of the main high geopotential center there is a large diurnal variation in wind speed and direction (compare Figs. 8a and 8c or Figs. 8b and 8d). The model results also indicate that the diurnal variation on the west and southwest sides of the main high centers are relatively small. Similar features can be noted in the observational data (Fig. 4).

The model derived divergence pattern associated with the diurnal forcing reveals a widespread influence of the oscillating mass source. At $t = 0$ (Fig. 9a), there is upper level convergence in the source region (i.e., subsidence at 500 mb according to the interpretation of the single internal vertical mode). The maximum upward motion is attained near the time when the forcing reaches its peak value (at $t = \tau/4$, Fig. 9b). At the end of the mass forcing, $t = \tau/2$, the vertical motion is still intense (Fig. 9c), switching to subsidence at $t = 3\tau/4$ (Fig. 9d). It is interesting to note that the implied sinking motion at this stage is about 60 percent in amplitude of the upward motion at $t = \tau/4$. Thus, a large response in the vertical motion is found at the source region after the forcing has been reduced to zero. Emanating out of the forcing region are cells which are associated with the patterns in the wind and geopotential fields (Fig. 8). Of particular interest is the magnitude of the divergence cells which form in the neighborhood of the source region; they tend to be 40 percent to 60 percent of the maximum value attained at the center of the source. In accordance with the ray path theory, the divergence centers are stronger near the equator, thus implying that the vertical motion response is smaller to the south of the forcing, as can be seen in Figs. 9b and 9d.

The similarity between Fig. 9b and the corresponding observed perturbation vertical motion (Fig. 3d) is an indication that the observed features are related to the pulsating tropical heat source. The model divergence to the east of the source is present in the observed perturbation ω -field as well as the zonally oriented bands of divergence/convergence in the Northern Hemisphere. Model runs with filtered Kelvin modes (not shown) indicate that the eastward propagating cells are basically determined by these fast modes.

The vorticity pattern lags the divergence at the forcing region, as shown in Fig. 10. The maximum vorticity

perturbation (positive/anticyclonic) is attained after the maximum convective forcing and the propagating vorticity cells are of much smaller amplitude than the main positive center and its negative envelope at the eastern sector. The quasi-stationary patterns are associated with the rotational modes which are slowly dispersive and the propagating cells are a manifestation of the gravity modes. Since gravity waves have much less vorticity than divergence, we observe small amplitude cells in spite of the large contribution by the fast modes to the total energy (Fig. 7).

Figure 11 shows the nondimensional divergence and vorticity at approximately the center of the mass source as a function of time. The divergence has a stronger diurnal signal compared to the vorticity and after the third cycle both have attained an approximately stationary regime. These results are representative of the area where the forcing is strong (inside the radius of e -folding damping). One interesting aspect of Fig. 11 refers to the lag between vorticity and divergence. Assuming that the maximum heat source occurs at approximately 18 LT (22 UTC), the maximum perturbation divergence is attained at 21 LT (01 UTC) and the corresponding minimum at 09 LT (13 UTC). Thus, the conventional upper air synoptic observations taken at 00 and 12 UTC coincide approximately with the maximum divergence signal. The same is not true for the vorticity, which is maximized between synoptic times and thus is not expected to show a strong signal in the observations, provided the maximum forcing occurs at the specified time. This fact probably explains the low statistical significance of the observed vorticity perturbation, as defined in section 2.

5. Discussion

Previous model results on the large scale response to the diurnal forcing on an f -plane (Paegle, 1978) in-

FIG. 11. Nondimensional divergence and vorticity at approximately the center of the mass source as a function of time.

dicate the rapid adjustment of the divergent component of the wind to the mass forcing. The fast divergent adjustment is not distorted by the incorporation of the β -plane approximation, which also allows for the dispersion of inertial gravity waves as well as the slow components. Our results also indicate that the dispersive character of the waves explains the temporal variations of the configurations that emanate out of the source region. The trapping of energy in the equatorial region is a consequence of this phenomena as well as the westward stretching of the rotational component associated with the upper anticyclonic cell.

The observed 500 mb vertical motion over the South American region and adjoining equatorial areas shows some agreement with the model derived divergence. The 00 UTC perturbation vertical motion is obtained a few hours after the maximum convection in the Amazon region (20 LT) as shown in the satellite derived phase of the diurnal component of the precipitation (Fig. 2c). Thus, the 12 UTC (08 LT) perturbation field is representative of the situation when the convective activity is a minimum over the area. The observed large amplitude of the diurnal cycle over the Amazon region seems to be well correlated with the large diurnal changes of the vertical motion in the adjoining regions as well as over large stretches along the equatorial region (Fig. 3). Figure 9 shows that the model divergence to the east of the forcing is also present in the observed ω_6 -field (Fig. 3). The model results also indicate that this feature is mostly related to the contribution of the eastward dispersion of the Kelvin waves although some eastward propagating waves also contribute to the southward displacement of the perturbation. The zonally oriented bands of divergence/convergence in the Northern Hemisphere are also present in the observations.

The model implied divergence/convergence patterns that emanate out of the source region have large amplitude in comparison with the local response (Fig. 9). Thus, the strong observed signal in the 500 mb vertical motion field perturbation patterns (Fig. 3), can be ascribed to the pulsating heat source over the Amazon region. It can also be argued that the rather remote divergent response at 3000 to 5000 km in the equatorial region in the model results is indicative of the impact of the Amazon heat source as a modulator of convection over the Atlantic intertropical convergence zone (ITCZ) and the eastern Pacific ITCZ. The phase of the diurnal precipitation cycle over the Atlantic ITCZ at about 20°W (maximum at 06 LT, Fig. 2c) fits very well with the model predicted upper divergence (Fig. 9). In addition, the Atlantic ITCZ should also be affected by the diurnal cycle over Africa, in accordance with the model divergence patterns that emanate to the west of the heat source. In fact, Minnis and Harrison (1984) satellite inferred diurnal variation of cloudiness indicates a coherent pattern, compatible with the divergent circulation suggested by the model

results. They suggest that the upper divergence over the Amazon undergoes a diurnal cycle that is not entirely compensated locally. The results indicate wave-like structures in deep cloudiness, emanating out of the Amazon region. This dynamical response to the diurnal forcing over South America (and similarly over equatorial Africa) is an alternative explanation to that proposed by Gray and Jacobson (1977) for the observed diurnal cycle in precipitation in the equatorial Atlantic.

The regular synoptic observing times of 00 and 12 UTC approximately correspond to the times of maximum and minimum convective activity, respectively, in the Amazon Basin (Fig. 2c). Due to the fast response of the divergent component to the forcing, the divergent signal should be a maximum in the analyses made at these times. In fact, this feature is evident in the observational results presented in Fig. 3, which shows the statistical significance of the diurnal signal in the vertical motion field. The model results also indicate that, considering the time scale of the internal mode for $c = 50$ ms, the extreme values of vorticity lag the forcing by about 6 hours (Fig. 11). Thus, it is not surprising that the vorticity field does not show a statistical significant signal in the observations.

There are some discrepancies between the observed diurnal cycle in the 500 mb level perturbation vertical motion field and the model derived upper divergence. The observed strong signal over northern Chile and the eastern Pacific that extends to the Atlantic Ocean is not reproduced in the model results. This can be partially related to the Andes effect which is not included in the model but it could also be related to nonlinear effects due to the presence of a strong subtropical jet at this latitude (Fig. 4). Finally, it should be mentioned that diurnal variation studies based on operational analyses are subjected to data assimilation problems which might favor one time over another (Jan Paegle, personal communication, 1983). This last comment points to the need of observational work based on data independent of the current assimilation techniques.

Acknowledgments. The authors benefited from discussions with Akira Kasahara, Wayne Schubert, Phillip Arkin and Patrick Minnis. We would also like to thank Dr. Greg Holland for reading the manuscript and for making several useful suggestions. Research support was provided by a Finep Grant IAG-USP. This research was also made possible through NOAA Grant NA84AAG01214 which provided Cray time at NCAR and support for the first author's visit to NCAR. Partial support was also provided by the NSF International Program NSF/INT8312513.

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