Latent Heat Release as a Possible Forcing Mechanism for Atmospheric Tides

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ABSTRACT

The consequences of the hypothesis of Lindzen (1978) that latent heat release may be a significant excitation mechanism for the semidiurnal atmospheric tide are examined in some detail. Harmonic analysis of hourly rainfall data from 79 tropical stations shows that the semidiurnal variation of rainfall in the tropics is $\sim$1 mm day^{-1} and has a phase near 0300 LST, just as Lindzen's theory requires. Analysis of data at 85 midlatitude stations shows that the semidiurnal rainfall oscillation there has its phase rather later (about 0600). The results of simple classical tidal theory calculations which indicate that the geographical distribution of the surface pressure response to latent heat forcing largely follows that of the forcing itself are presented. This result is then used to suggest a plausible explanation for the observed seasonal cycle of the semidiurnal pressure oscillation in midlatitudes. Further calculations show that the magnitude of the non-migrating components of the semidiurnal barometric oscillation produced by latent heat excitation is not likely to be unrealistically large. These calculations also suggest that Lindzen's hypothesis might be verified by observing the phase of the semidiurnal pressure oscillation in particularly arid regions.

The rainfall observations also show a strong diurnal (24 h) component in the rainfall both in the tropics and in midlatitudes. The effects of latent heat release on the 24 h tide are briefly discussed.

1. Introduction

In the 1960's it was shown that many of the observed features of the solar diurnal and semidiurnal atmospheric tides can be accounted for with the "classical" theory (i.e., by solving the inviscid linear forced wave problem for an atmosphere with no mean winds or meridional temperature gradient) if forcing from direct absorption of solar radiation by both ozone and water vapor is included (e.g., Butler and Small, 1963; Lindzen, 1966, 1967, 1968). Several discrepancies between the theory and observations remained, however. One of these involved the phase of the semidiurnal surface pressure oscillation [$S_2(p)$]. The inviscid, adiabatic theory predicts that the maxima in $S_2(p)$ should occur at roughly 0845 (and 2045) LST [see the results for the equatorial temperature profile in Lindzen (1968)] while the observed phase[1] equatorward of about 45° latitude is generally between 0930 and 1000 LST. Inclusion of viscosity (Lindzen and Blake, 1971) or inclusion of mean winds in the basic state (Lindzen and Hong, 1974) apparently tends to move the calculated phase close to 0900, but even with such effects a large phase difference between theory and observations still remains. Another discrepancy between the observations and the theory concerns the vertical structure of the semidiurnal horizontal wind field [$S_2(\mathbf{v})$]. The calculations predict that $S_2(\mathbf{v})$ should have a node at $\sim$28 km. Observations by Wallace and Tadd (1974) have shown that there is no node in $S_2(\mathbf{v})$ below 30 km in any season [although a node often appears at roughly 50 km (Reed, 1972)].

Recently Lindzen (1978) has proposed that latent heat release may be an important source of excitation for the semidiurnal tide and that this additional forcing mechanism may allow one to eliminate the discrepancies between the theory and observations of $S_2(p)$ and $S_2(\mathbf{v})$. His reasoning may be summarized as follows. The observed and calculated values of $S_2(p)$ at a typical station (located anywhere equatorward of 45° latitude) are shown on a harmonic dial in Fig. 1. Lindzen noted that the theory and observations could be reconciled if another excitation were included in the calculation which produced the surface pressure response indicated by the dashed line in Fig. 1. He was able to show that a semidiurnal latent heat source extending throughout the lower and middle troposphere with an amplitude of $\sim$1.2 mm day^{-1} (liquid water equivalent) at the equator (and modulated in some reasonable manner in the meridional direction) and a phase of roughly 0330 LST indeed would produce the required response when employed in the classical tidal theory calculations. Lindzen then Fourier analyzed hourly rainfall records at several tropical

[1] It should be noted that the term "phase" in this paper refers to the local time at which the diurnal or semidiurnal Fourier component of the variation of a particular quantity (e.g., pressure) reaches its maximum.

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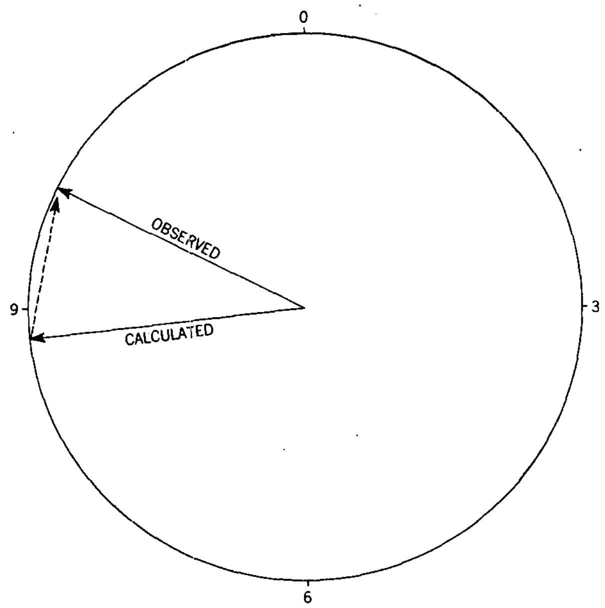

FIG. 1. Harmonic dial representation of the observed value of $S_2(p)$ and the value calculated using classical tidal theory with ozone and water vapor excitation. At the equator the amplitude of both arrows would be ~1.2 mb.

stations and found that the phase of the semidiurnal rainfall oscillation [$S_2(R)$] was in most cases quite close to 0330 and that the amplitude was generally ~1 mm day^{-1}. From this he concluded that latent heat forcing was a plausible explanation for the phase discrepancy in $S_2(p)$. He also noted that the amplitude of the diurnal rainfall oscillation [$S_1(R)$] is in most cases smaller than that of $S_2(R)$ and that the phase of $S_1(R)$, when expressed in local time, was quite variable from station to station. From these observations he speculated that latent heat release is not a significant forcing mechanism for the 24 h tide. Lindzen was further able to show that a strong semidiurnal source in the troposphere would eliminate the node at 28 km in the calculations of $S_2(v)$.

The purpose of the present work is to examine Lindzen's hypothesis in some detail. In Section 2 the results of Fourier analysis of a large quantity of hourly rainfall data will be presented. The remainder of the paper is devoted to a consideration of the geographical and seasonal dependence of the tidal fields that might be produced by latent heat excitation. In Section 3 some general remarks will be made concerning the form of the surface pressure response that might result from tropospheric tidal forcing. In Section 4 the role of latent heat forcing in the seasonal variation of $S_2(p)$ will be examined. Section 5 is a study of the "non-migrating components" (see Chapman and Lindzen, 1970) that are produced when latent heat release is included in theoretical calculations of $S_2(p)$. In Section 6 the effects of latent heat on the diurnal (24 h) atmospheric tide are discussed. Finally, the conclusions of the present work are summarized in Section 7.

It might be appropriate to point out some matters that are not discussed in this paper. First of all, no attempt is made to determine the cause of the latent heat excitation itself. Discussions of the problem of diurnal variations of moist convection have been given by Malkus (1964), Brier and Simpson (1969), Gray and Jacobson (1977) and Lindzen (1978). Suffice it to say that nothing approaching a complete theory exists. Another important point is that two assumptions implicit in Lindzen's work will also be used in the present paper, i.e., that rainfall is a good (and almost instantaneous) indicator of the latent heat release above a particular location, and that the latent heat excitation may be used in a simple manner as a heat source in the large-scale dynamical equations. This last assumption, of course, is a traditional one in tropical meteorology, but it is worth noting that it is usually applied to disturbances with periods much longer than those of present interest.

2. Analysis of rainfall data

Lindzen (1978) performed Fourier analysis on hourly rainfall records from about 20 tropical island stations which were obtained from Brier and Simpson (1969), Finkelstein (1964), Inchauspe (1970), Jacobson (1976) and Ray (1928). In this section the results of a similar analysis of another 179 stations will be presented. The raw data for this study were taken from Alexander (1938), Armstrong (1934), Bennet (1929), Besson (1930), Carter (1924), Cook (1939), Cox and Armington (1914), Dhar (1960), Feldwisch (1924), Flora (1924), French (1924), Hall (1906), Hallenbeck (1917), Hann (1906), Hastenrath (1970), Iyer and Zafar (1946), Jin-bee (1959), Johnson (1976), Kincer (1916), Kobayashi (1893), Lessmann (1968), Maunder (1957), Mulky (1958), Narasimham and Zafar (1947), Nunn (1922), Prasad (1970, 1974), Ramage (1952, 1964), Ray (1925, 1927), Sanford (1923), Schumann (1955), Shipman (1925), Snow (1976), Steinhauser (1965), Thompson (1965), Voorhees (1928), Ward (1908) and Watts (1955). An indication of the resulting station distribution is given in Table 1. It should be noted that, in contrast to Lindzen's study, most of the 179 stations are not from small islands but are generally what could be termed coastal or inland stations.

The data at each station consisted of published values of hourly (112 stations), two-hourly (50 stations) or three-hourly (17 stations) rainfall rates throughout the day which had been averaged over several years. For the purposes of the present study, each station was classified as being either tropical (equatorward of 30° latitude) or midlatitude (poleward of 30°). For 89 of the 94 tropical stations and 33 of the 85 midlatitude stations, data were available for the entire year. The data for the remaining stations covered only some fraction of the year (generally the summer months). The average record length of all the stations used in this study was over 11 years, although there were about 20 stations with record lengths of only two or three years. The inclusion of these stations is not meant to indicate that the diurnal and semidiurnal rainfall oscillations can be reliably determined from such short records. Indeed, as the discussion in Appendix A shows, it seems that considerably longer records should be used. However, in view of the method of analysis and presentation used in this study and because of the scarcity of published hourly rainfall records, it was decided to include all available stations with records of at least 24 months.

The data were Fourier analyzed in a straightforward manner to determine the harmonic coefficients for the diurnal and semidiurnal rainfall oscillations, i.e.,

$$\left.\begin{aligned} A_n &= (2/N) \sum_{q=1}^{N} R_q \cos(2\pi qn/24), \quad n = 1, 2 \\ B_n &= (2/N) \sum_{q=1}^{N} R_q \sin(2\pi qn/24), \quad n = 1, 2 \end{aligned}\right\},$$

where R_q is the average rainfall rate at $t = 24q/N$ h, and where $N = 24$, 12, 8 for hourly, two-hourly and three-hourly data, respectively. From the A_n and B_n it is easy to determine the amplitude and phase of $S_n(R)$. It should be noted that the phases used in the present observational study are referred in most cases to local standard time, which may, of course, differ slightly from local solar time. A complete list of all 179 stations with

TABLE 1. The geographical distribution of the stations used for the analysis of Section 2.

Region	Total number of stations	Stations with annual mean data	Number of tropical stations
United States	17	11	1
Europe	50	6	—
Japan	10	10	—
New Zealand	4	4	—
Southern Africa	18	18	13
East Africa	43	42	43
India	18	17	18
Malaya, Hong Kong	4	1	4
Central America	7	7	7
South America	7	7	7
Carribean	1	1	1
Total	179	124	94

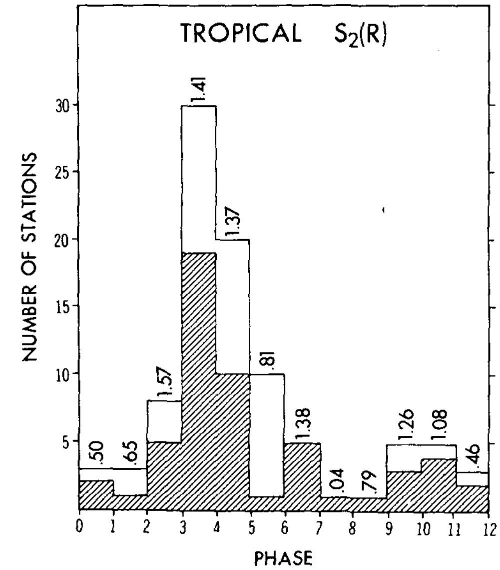

FIG. 2. Histogram showing the number of stations with observed phase of the semidiurnal rainfall oscillation between the hours indicated. The numbers above each bar of the histogram give the average amplitudes of $S_2(R)$ (mm day^{-1}) for the stations with the indicated phase. This graph is for 94 stations between 30°S and 30°N. The shaded histogram is obtained if only those stations with at least 120 months of data are employed.

the calculated amplitudes and phases of $S_1(R)$ and $S_2(R)$ is available from the author on request.

Fig. 2 is a histogram of the number of tropical stations for which the phase of $S_2(R)$ lay between the hours shown. The numbers above each bar of the histogram give the average amplitude (mm day^{-1}) of $S_2(R)$ for those stations with the indicated phase. The shaded histogram is obtained if only stations with at least ten years record are used. The most obvious feature in this figure is the large peak around 0330. This confirms Lindzen's observations in a rather striking manner and shows that they apply to coastal and inland stations as well as to islands. The average amplitude of $S_2(R)$ for the stations in this peak is greater than 1 mm day^{-1}. Since 43 of the 94 tropical stations used in this study are located in eastern Africa, it is worth noting that if a similar histogram is prepared without including the eastern African stations, then the general features remain unchanged from those of Fig. 2.

Fig. 3 is similar to Fig. 2 but is for the 85 midlatitude stations. Once again there is a large peak in the graph but it is broader than in the tropical case and it is centered at roughly 0600 LST. The average amplitude of $S_2(R)$ for stations in this peak is ~0.4 mm day^{-1}. When Fig. 3 is redrawn without employing the 50 European stations its qualitative appearance remains the same.

Figs. 4 and 5 are similar histograms for the diurnal rainfall oscillation. For both the tropical

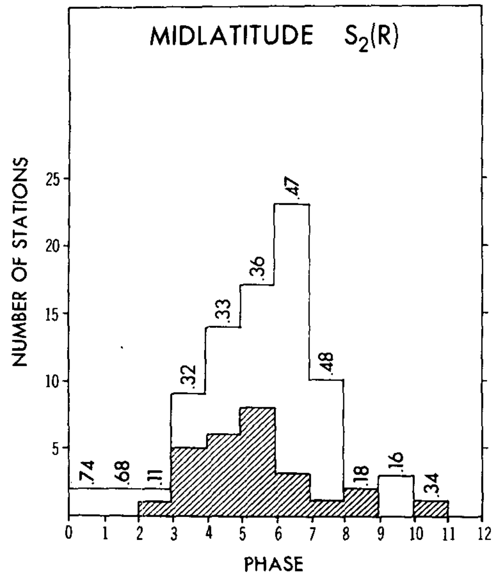

FIG. 3. As in Fig. 2 except for 85 stations poleward of 30° latitude.

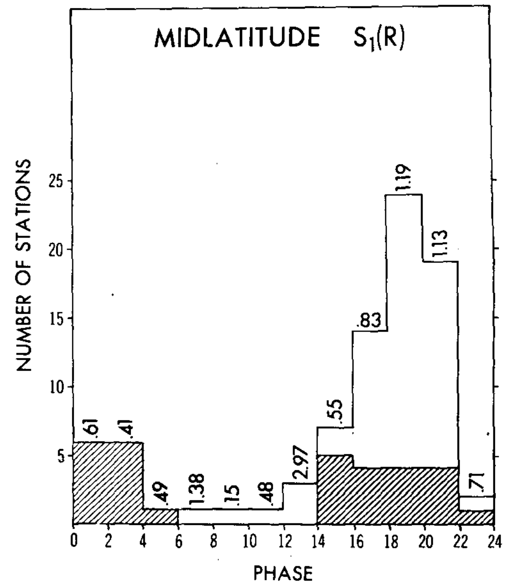

FIG. 5. As in Fig. 3 except for the diurnal rainfall oscillation.

and midlatitude stations there is a strong tendency for $S_1(R)$ to have its phase in the late afternoon or early evening. The average amplitude of $S_1(R)$ is considerably larger than that of $S_2(R)$. Both of these features contrast with Lindzen's rainfall observations. The difference is presumably to be explained by the fact that Lindzen's data come from small tropical islands while most of the data used in this study come from localities with more continental meteorological conditions. From these observations, one is tempted to speculate that the rainfall regime is mainly semidiurnal over oceans and small islands but has a strong diurnal component due to afternoon convection over the major land areas. The consequences of this speculation for the theory of the diurnal atmospheric tide will be examined in Section 6.

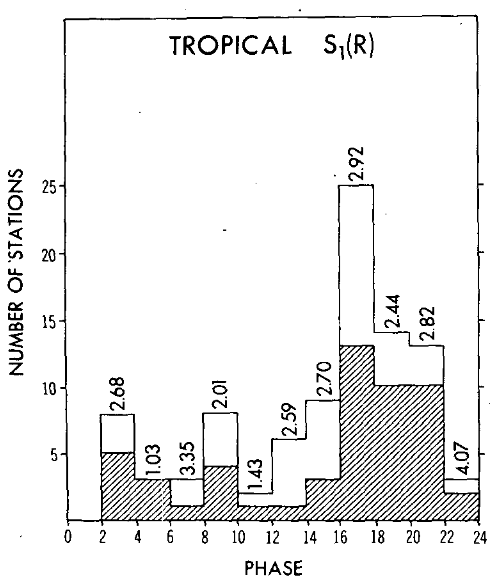

FIG. 4. As in Fig. 2 except for the diurnal rainfall oscillation.

Features of $S_1(R)$ and $S_2(R)$ that are of interest for the tidal theorist are their seasonal variation, particularly in midlatitudes (see Section 4). Although seasonally and zonally averaged total precipitation varies negligibly throughout the four seasons at all latitudes (Newell *et al.*, 1974), winter precipitation at a midlatitude site might be expected to undergo smaller diurnal and semidiurnal fluctuations than summer precipitation, which is presumably more influenced by such factors as surface heating and cloud-top cooling. Unfortunately, it is impossible to conclusively prove or disprove this speculation on the basis of the data used in this study. In the present data set there are only nine midlatitude stations with record lengths of at least five years and with records divided by season. The ratios of the amplitudes of the $S_n(R)$ in summer to those in the winter have been calculated for each of the stations and the result is shown in Table 2. At all nine stations the ratio for $S_1(R)$ is larger than one, while for $S_2(R)$ the ratio is greater than one at seven of the stations. If these few observations turn out to be representative of the midlatitude region as a whole, then one can certainly expect the diurnal and semidiurnal latent heat excitation to be considerably smaller in winter than in summer.

3. Modeling semidiurnal tropospheric forcing

Lindzen (1968) has investigated the response to a thick (~60 km) semidiurnal forcing in the stratosphere

TABLE 2. The ratio of the amplitude of $S_1(R)$ in summer to that in winter [ratio(1)] and the same ratio for $S_2(R)$ [ratio(2)]. Summer is defined as July–September for the Northern Hemisphere stations and as January–March for the Southern Hemisphere stations. Winter is defined as January–March in the Northern Hemisphere and July–September in the Southern Hemisphere.

Station	Reference	Latitude	Longitude	Ratio(1)	Ratio(2)
Fort Smith	Shipman (1925)	35.4°N	94.4°W	3.5	20.2
Nashville	Nunn (1924)	36.1°N	86.7°W	2.67	4.1
Oklahoma City	Alexander (1938)	35.4°N	97.6°W	9.0	21.5
Topeka	Flora (1924)	39.0°N	91.7°W	4.8	1.84
Tokyo	Kobayashi (1893)	35.7°N	139.8°E	1.29	3.3
Christchurch	Maunder (1957)	43.5°S	172.6°E	1.84	1.43
Hokitika	Maunder (1957)	42.7°S	170.9°E	5.3	0.86
Kelburn	Maunder (1957)	41.3°S	174.8°E	4.1	1.33
Ruakura	Maunder (1957)	37.7°S	175.3°E	2.23	0.60

and mesosphere (i.e., the O_3 shortwave heating). He found that the surface pressure response should be relatively unaffected by the details of the meridional distribution of the excitation. This is a consequence of the fact that the gravest Hough mode which represents a semidiurnal wave following the sun (called the $\Theta_{2,2}$ mode) has a vertical wavelength of ~200 km, while all the other modes are associated with much shorter vertical wavelengths (<50 km) and thus only $\Theta_{2,2}$ is efficiently excited by a thick source. There is no particular reason, however, to believe that the $\Theta_{2,2}$ mode must be such an important component of the response to a much thinner source located near the ground. In this section, it will be argued that the surface pressure response to tropospheric tidal forcing is not necessarily dominated by $\Theta_{2,2}$ but, in fact, largely follows the geographical details of the forcing.

To see intuitively why this might be so, one can imagine a tidal forcing confined to a narrow latitude band. Since atmospheric tides are internal gravity waves which are somewhat modified by rotation and sphericity, one might hope that the response to this forcing could be deduced (at least approximately) from simple gravity wave theory. It is well known that the response to an isolated, monochromatic source of gravity waves is largest along the characteristic lines radiating out from the source region. These characteristic lines have slopes equal to the angular frequency of the waves divided by the Brunt-Väisälä frequency. For the semidiurnal tide in the troposphere this slope is typically 10^{-2}, so that if the top of the forcing region were located at height H, one might expect that the response would be reasonably well localized within a latitude band of width 200 H. For an H of 10 km, this amounts to about 20° of latitude. It should be noted that this sort of argument would not be applicable to a source that extended throughout stratosphere, since the wave characteristics would traverse a significant fraction of the globe before reaching the ground and, thus, the effects of sphericity and rotation would presumably become extremely important.

In order to demonstrate that this geographical "localization" of the surface pressure response actually should occur under realistic conditions, some classical tidal theory calculations were performed.[2] The basic-state temperature profile and the vertical distribution of the forcing [denoted by $J(z)$ and defined as the thermotidal heating per unit mass per unit time] are shown in Figs. 6 and 7, respectively. The J chosen is meant to be typical of the latent heat release in cumulus convection and is almost identical to that used by Lindzen (1978), where it was given some justification in terms of Arakawa's cumulus parameterization scheme (Arakawa and Schubert, 1974).[3] The solid curve in Fig. 8 shows the amplitude of the surface pressure response to a semidiurnal latent heat source modulated in the meridional direction by a slit function 20° in width and centered at the equator, as calculated with adiabatic, inviscid theory. The response is somewhat broader than the forcing, but it is certainly narrower than $\Theta_{2,2}$ (which has a half-width of roughly 60°). Fig. 9 shows the amplitude of the response to a slit function centered at 40°N. Once again, the response is local. In both cases, the calculated surface pressure response is very nearly in phase with the forcing, i.e., the maximum in heating rate leads the maximum in surface pressure by 3 h (at least in regions where the amplitude of the response is appreciable).

It seems reasonable to ask how these results might be changed if dissipation were included in the

[2] In the classical tidal theory calculations discussed in this section and in the next section, the meridional distributions of the excitations were expanded in the 14 gravest Hough functions associated with westward traveling wavenumber 2. For each Hough function the vertical structure equation was solved numerically using the algorithm described in Chapman and Lindzen (1970). In the integration of the vertical structure equation 2000 levels between $z = 0$ and $z = 179$ km were employed and a radiation condition was imposed at the top level.

[3] The sharp cutoff in $J(z)$, of course, is unrealistic. However, it is interesting to note that the $J(z)$ in Fig. 7 actually does resemble the *total* latent heating profile at the equator as simulated in a general circulation model by Manabe *et al.* (1974).

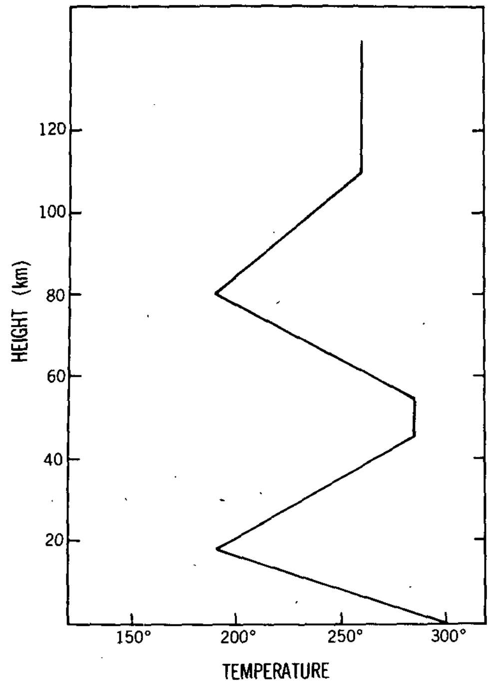

FIG. 6. The basic-state temperature profile used in the classical tidal theory calculations.

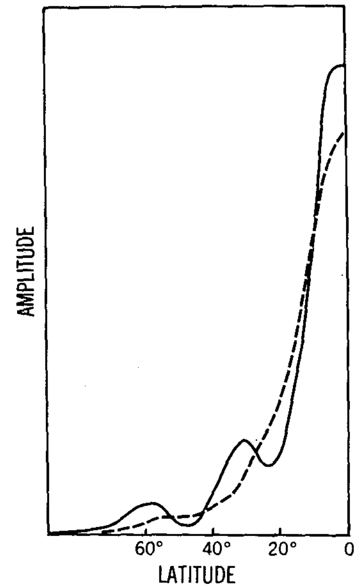

FIG. 8. The amplitude of the surface pressure response to a semidiurnal latent heat source confined between 10°S and 10°N. The solid curve shows the results of a calculation employing inviscid, adiabatic tidal theory, while the dashed curve shows the results when Newtonian cooling is included in the calculation. The amplitude scale (and the relative amplitudes of the solid and dashed curves) is arbitrary.

tidal calculations, particularly as the modal decomposition of such localized forcings have significant contributions from high-order Hough functions. The dotted curve in Fig. 8 shows the result of repeating the calculation for the equatorial slit function with Newtonian cooling included in the vertical structure equation. The Newtonian cooling coefficient α above 30 km was taken from Dickinson (1973). Below 18 km, α was taken to be zero, and between 18 and 30 km α was linearly interpolated. The results show that even with thermal damping

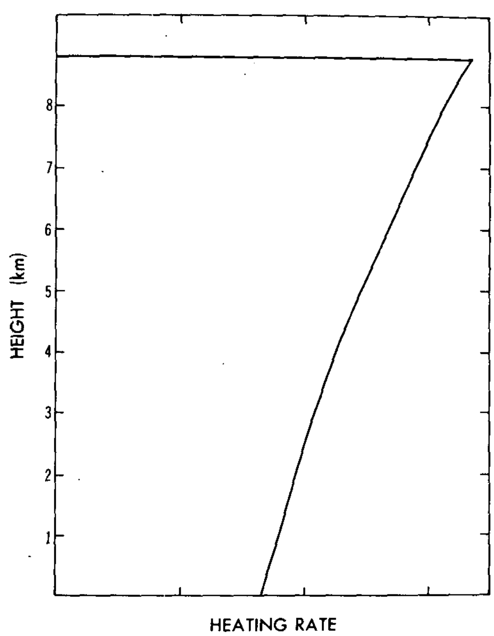

FIG. 7. The vertical distribution of the latent heat forcing $J(z)$ used in the tidal calculations. The horizontal scale is arbitrary.

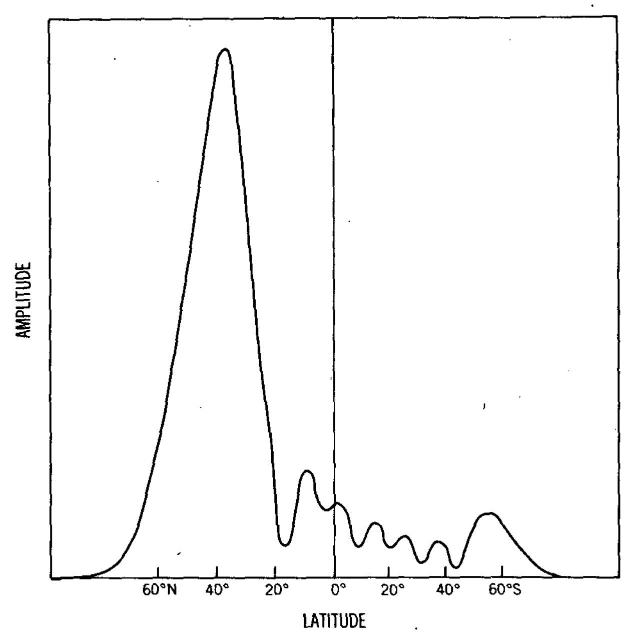

FIG. 9. The amplitude of the surface pressure response to a semidiurnal latent heat excitation confined between 30 and 50°N, as calculated using inviscid, adiabatic theory. The amplitude scale is arbitrary.

the surface pressure response is quite well localized to the forcing region.

The question of whether mechanical dissipation can affect the results of the calculations discussed above is much harder to answer, since viscosity cannot be incorporated in separable tidal theory calculations. However, some idea of the magnitude of the effect may be obtained from consideration of the theory of plane gravity waves propagating away from a source region in a viscous, unbounded atmosphere. Below the source region, the amplitude of the pressure perturbation (normalized by the square root of the basic state density) would be expected to decay roughly as $\exp(\beta z/\lambda)$, where λ is the vertical wavelength and where $\beta = (2\pi)^2\nu/\sigma\lambda^2$, where ν is the kinematic eddy viscosity and σ the angular frequency of the wave. Now in the calculations presented in Figs. 8 and 9, the shortest vertical wavelengths associated with the Hough functions needed to resolve the forcing were ~5 km. For such modes $\beta \approx 0.1$ if a value of ν of 10 m^2 s^{-1} is used. Since the distance from the forcing region to the ground is about one wavelength, one expects the wave amplitude to be reduced by only ~10% at the ground. Thus it seems reasonable to expect the results of the present calculations to be relatively unaffected by mechanical dissipation.

It is interesting to consider the consequences of the localization of the surface pressure response to the forcing region in light of the results discussed in Section 2. A little thought shows that, in order to eliminate the discrepancy between the observed and calculated phases of $S_2(p)$ in both the tropics and midlatitudes, the response represented by the dotted line in Fig. 1 must have the same phase at each latitude and its amplitude should be modulated in the meridional direction in the same manner as the response to the ozone and water vapor excitations. In so far as the localization holds, this means that the latent heat source itself must have the same phase (around 0330) in both the tropics and midlatitudes. If the results presented in Section 2 are truly representative, then there appears to be a serious problem since the rainfall analysis indicated that in midlatitudes the latent heat excitation has its phase around 0600. This would lead to a surface pressure response with a phase of roughly 0300 which would, of course, have a negligible effect on the computed phase of $S_2(p)$ in midlatitudes.

4. Modeling the seasonal variation of $S_2(p)$

The only investigation of the seasonal variation of the solar semidiurnal atmospheric tide appears to be that of Lindzen and Hong (1974), who devote very little discussion to their surface pressure results. In fact, $S_2(p)$ has a quite remarkable annual cycle in the midlatitude regions.

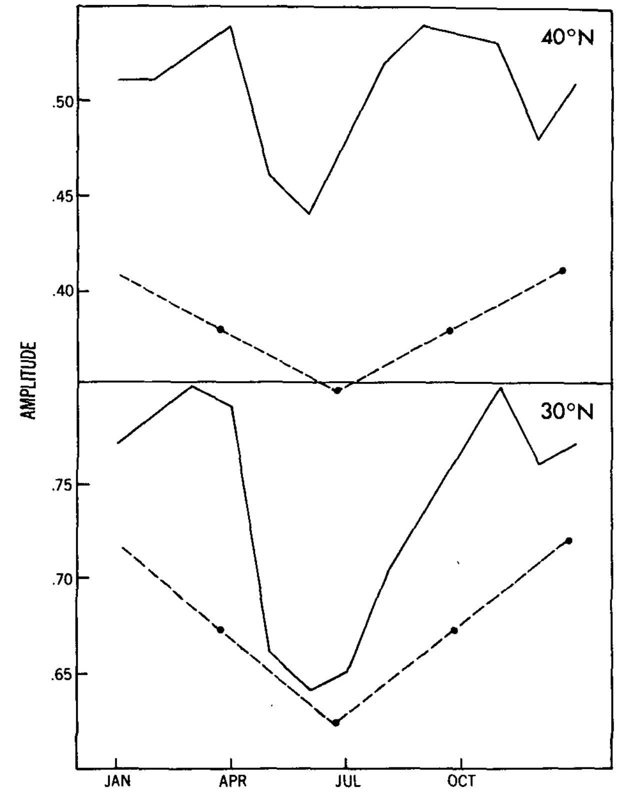

FIG. 10. The solid curve in the upper figure shows the observed amplitude of $S_2(p)$ for American stations between 35 and 45°N. The solid curve of the lower figure is for stations between 25 and 35°N. The dots in the two figures show the theoretically calculated values for the solstices and the equinox at 30 and 40°N. The dots have been connected with the dashed lines. The solid curves are taken from Spar (1952).

The solid curves in Figs. 10 and 11 show the observed monthly amplitudes and phases of $S_2(p)$ averaged over 39 American stations in the latitude belt from 35 to 45°N and for 26 stations between 25 and 35°N as reported by Spar (1952). From Fig. 10 it can be seen that the amplitude of $S_2(p)$ in midlatitudes undergoes a seasonal cycle with prominent annual and semiannual components. The primary minimum occurs in June and a secondary minimum appears in December. From Fig. 11 it is apparent that the phase is a minimum in January and a maximum in June, with a 40–50 min difference between these extreme values.

In order to see how well classical tidal theory with just the traditional H_2O and O_3 excitations can account for these seasonal effects, the amplitude and phase of $S_2(p)$ at the solstices and at the equinox were calculated. The O_3 and H_2O forcing was taken from Lindzen and Hong (1974). All other aspects of the calculations were the same as in Section 3. The results for 40 and 30°N are shown by the dashed lines in Figs. 10 and 11.

The theoretical results are deficient in several respects. The calculated amplitudes are all slightly

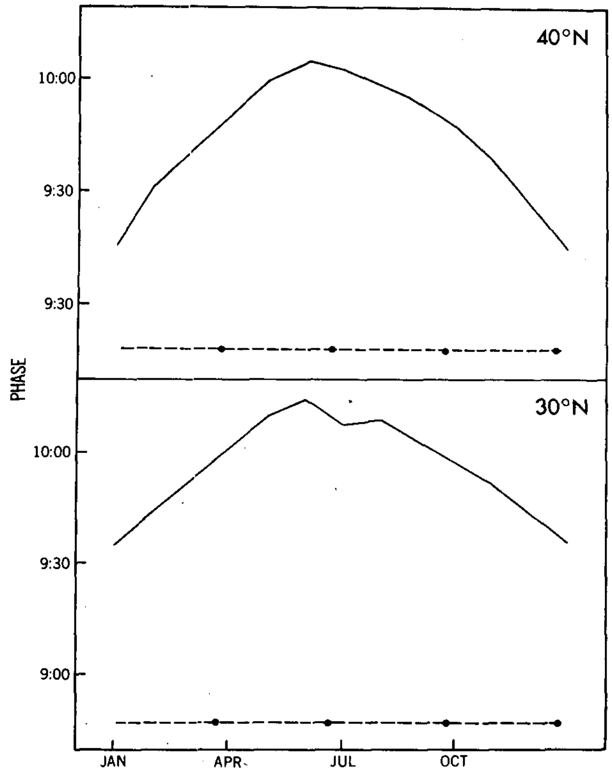

FIG. 11. As in Fig. 10 except for the phase.

smaller than the observed ones (which may possibly reflect a nonuniform distribution of Spar's stations throughout the latitude belts considered). The seasonal variation of the amplitude in the calculation is also smaller than in the observations, with the semiannual component being noticeably suppressed in the theoretical results. This may be due to the simplified form of the excitation used. In particular, the part of Lindzen and Hong's excitation that is symmetric about the equator undergoes no seasonal cycle. More recent calculations of the O_3 forcing by Forbes and Garrett (1978) show that the amplitude of the $\Theta_{2,2}$ component of the excitation is somewhat smaller at the solstices than at the equinox.

While there may be some problems in reconciling the observations and calculations of the amplitude of $S_2(p)$, the situation with the phase is much worse. As noted before, the annual mean calculated phase is roughly an hour behind that which is observed. In addition, the calculated phase has almost no seasonal cycle whatsoever. This is because the surface pressure response in each season is dominated by the $\Theta_{2,2}$ mode and, if only O_3 and H_2O forcing is included, the phase of the excitation of $\Theta_{2,2}$ does not vary throughout the year. Thus, it would seem that the theoretical results for the phase are not likely to be sensitive to the details of the O_3 and H_2O forcing used.

If, for the moment, one assumes that the rainfall results of Section 2 are not representative and that the latent heat forcing in midlatitudes, in fact, does have the required 0330 phase, then it is of interest to ask if Lindzen's hypothesis can be used to account for the large seasonal cycle of the phase of $S_2(p)$. Given that the phase of the contribution to $S_2(p)$ from O_3 and H_2O forcing remains constant throughout the year, the observed seasonal cycle of the phase can be explained simply by postulating that the amplitude of the contribution to $S_2(p)$ from latent heat release undergoes a large annual cycle in midlatitudes with a minimum in the winter and a maximum in the summer. From the discussion of the previous section, this would imply that the latent heat forcing in the midlatitude region must have a similar annual cycle. This, in turn, is consistent with the hypothesis presented in Section 2 that diurnal effects on moist convection may be considerably weaker in winter than in summer. Thus Lindzen's work can be used as the basis for a plausible explanation for the seasonal variation of the phase of $S_2(p)$, a phenomenon which would seem to be very difficult to explain in any other reasonable manner.

5. Geographical variations in $S_2(p)$

Kertz (1956) and Haurwitz and Cowley (1973) have expanded the observed values of $S_2(p)$ throughout the world into a series of traveling waves. They wrote

$$p(\theta,\phi,t) = \sum_{m=-10}^{14} g_m(\theta) \exp[i(m\phi + \sigma t_u)], \quad (1)$$

where t_u is Greenwich Mean Time and where ϕ is the longitude with east taken as positive. The main result of this analysis is that the $m = 2$ component (i.e., the component that exactly follows the sun) is an order of magnitude larger than any of the other components (which will be referred to, somewhat inexactly, as non-migrating components). The dominance of the $m = 2$ component is such that the amplitude and local time phase of $S_2(p)$ are observed to have only small deviations from zonal symmetry, at least equatorward of 45° latitude. Since shortwave absorption by a zonally symmetric distribution of absorbers would produce only an $m = 2$ component, these surface pressure observations are easy to understand if the only important source of excitation for S_2 is O_3 and H_2O heating. However, if latent heat release is also a significant forcing mechanism for the semidiurnal tide, then the smallness of the non-migrating components seems much more difficult to understand given the large geographical variations which are observed in the total precipitation throughout the world. In this section, an attempt will be made to

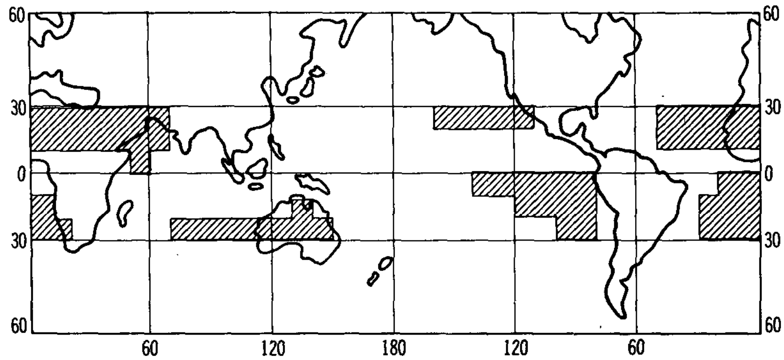

FIG. 12. The dry zones used in the tidal calculations of Section 5.

deduce some of the features of the geographical distribution of the surface pressure response which would be expected to result from latent heat tidal forcing.

Due to the lack of uniformity in the geographical distribution of stations for which observations of $S_2(R)$ exist (i.e., those stations used in Lindzen, 1978, and those used in Section 2 above), there seems to be no practical way of constructing a truly realistic model for the geographical distribution of the latent heat forcing, so the following simple procedure was adopted. A map of average total precipitation throughout the world was examined. It was reasoned that, in those areas where there is <500 mm year^{-1} rainfall, it is unlikely that the semidiurnal rainfall oscillation has an amplitude even roughly comparable to the 1 mm day^{-1} which Lindzen's theory requires in the tropics. Therefore, in the tropical regions that receive <500 mm year^{-1} rainfall the latent heat tidal forcing can be taken to be virtually zero for present purposes. Thus, a model latent heat forcing function was constructed by setting

$$J(\theta,\phi,z,t) = 0$$

in the "dry regions", and

$$J(\theta,\phi,z,t) = f(z)\Theta_{2,2}(\theta) \exp[i\sigma(t - 3)]$$

elsewhere. In these expressions, the dry regions are those areas equatorward of 30° latitude that get <500 mm year^{-1} average rainfall, $f(z)$ is taken from Fig. 7, and t is local solar time. The positions of the dry areas used in this study were determined from the rainfall data of Jaeger (1976) and are shown in Fig. 12. The meridional modulation of the latent heat forcing was chosen to be proportional to $\Theta_{2,2}$ so as to give good results for the traveling wave at all latitudes.

While the latent heat model employed in this study certainly cannot be considered realistic, it does possess two important features which will be used later in the analysis of tidal calculations. First of all, the model latent heat forcing over the dry regions probably approximates the actual forcing there quite well. Thus, if it can be shown that the response to the latent heat excitation is well localized within the forcing region, then the calculated $S_2(p)$ within the dry regions themselves should be quite realistic. The other point is that, since the forcing in the dry regions is realistic and outside the dry regions is taken to be very smooth (certainly smoother than the actual latent heat forcing), one would expect the model forcing to underestimate the magnitude of the non-migrating components (except, possibly at the very large wavenumbers which will be produced by the sharp edges of the dry zones in the model forcing).

Classical tidal theory calculations were performed using the model latent heat excitation. In these calculations J was expanded in 740 Hough functions representing the 20 gravest modes for each of wavenumbers $m = -16, -15, \ldots, 0, 1, 2, \ldots, 20$, where a minus sign indicates eastward-traveling waves. The vertical structure equation was then solved for each Hough mode.[4] Finally, the results were resynthesized to obtain the geographical distribution of the surface pressure oscillation. To this result was added another component of $S_2(p)$,

$$A\Theta_{2,2}(\theta) \exp\{i[\sigma(t - 8.75)]\},$$

meant to simulate the surface pressure oscillation excited by O_3 and H_2O shortwave absorption. The constant A was adjusted to give reasonable values of the zonally averaged phase in the tropics.

The calculated $S_2(p)$ field was analyzed into traveling waves at each latitude in the manner of Eq. (1). Then an average amplitude of each wave was defined by

$$G_m = \int_{-90°}^{90°} g_m(\theta)\cos(\theta)\,d\theta \left[\int_{-90°}^{90°} \cos(\theta)\,d\theta\right]^{-1}.$$

[4] The solution of the vertical structure equation was performed in the manner described in Section 3, except that only 1000 levels between $z = 0$ and $z = 140$ km were used.

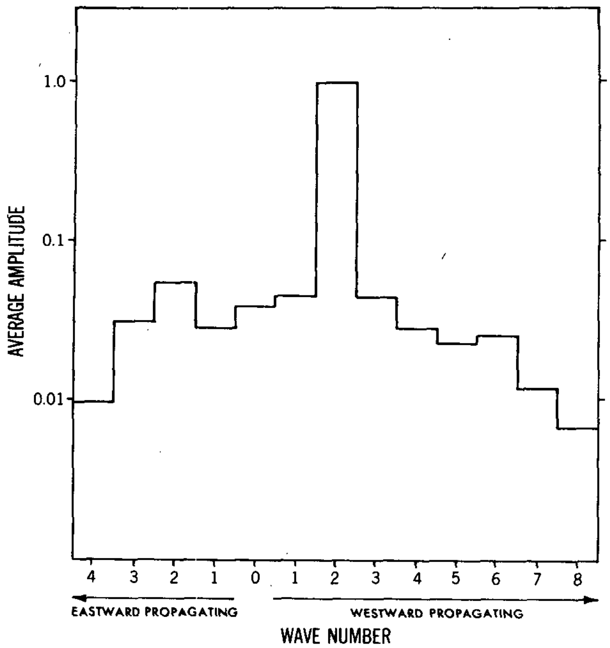

FIG. 13. The average amplitudes of the different zonal waves in the Fourier expansion of the $S_2(p)$ field calculated in Section 5. Note that the amplitudes are plotted on a logarithmic scale, and that G_2 has been taken to be one.

The G_m's are plotted in Fig. 13 on a scale where G_2 is equal to one. This figure can be compared with a similar one computed from observed values of $S_2(p)$ which is given by Haurwitz and Cowley (1973). While the details of Fig. 13 and the Haurwitz and Cowley spectrum are not in very good agreement, the magnitude of the largest non-migrating component in both cases is roughly a factor of 10 smaller than that of the main traveling wave ($m = 2$). Thus, it seems difficult to challenge Lindzen's hypothesis on the ground that latent heat excitation would produce non-migrating components that are larger than those which are actually observed (unless, of course, a more realistic latent heat excitation would have considerably larger non-migrating components than the simple model used here). This is an important point since prior to Lindzen's (1978) paper, the prime objection to latent heat release as an excitation mechanism for atmospheric tides was the large geographical variations expected in the response (e.g., Lindzen and Blake, 1971).

The amplitude contours for the calculated $S_2(p)$ field in the tropics are shown in Fig. 14. The deviations from zonal symmetry are hardly spectacular and have magnitudes quite comparable to those seen in the map of observed amplitude isopleths given by Haurwitz (1956). The two most conspicuous features in Fig. 14 are the maxima over the Indian Ocean and over the eastern Pacific. The first maximum is actually observed (e.g., Haurwitz, 1956), but observations in the tropical eastern Pacific area are too sparse to confirm or deny the existence of the second computed maximum. A final point that should be made about Fig. 14 is that the amplitude anomalies do not seem to be related in any simple way to the positions of the dry regions used in the calculations.

Fig. 15 shows the 0900, 0915 and 0930 LST phase lines for the calculated $S_2(p)$ field. It can be seen that early phase anomalies appear over most of the dry regions. This is easy to understand in terms of a localized response to the latent heat forcing. Over the dry areas one expects very little contribution to $S_2(p)$ from the latent heat excitation and thus the tidal oscillation there is primarily forced by shortwave absorption and has a phase close to 0900. As mentioned above, the most credible results for this calculation are those for the dry regions themselves. This suggests that one could verify Lindzen's hypothesis by simply observing the phase of $S_2(p)$ for stations in the dry areas. Unfortunately, there are very few determinations of $S_2(p)$ in the dry regions of the world (Haurwitz, 1956). It is of interest to survey what data do exist, however.

Of those stations listed in Haurwitz' paper and located in the North African dry zone, Cairo,

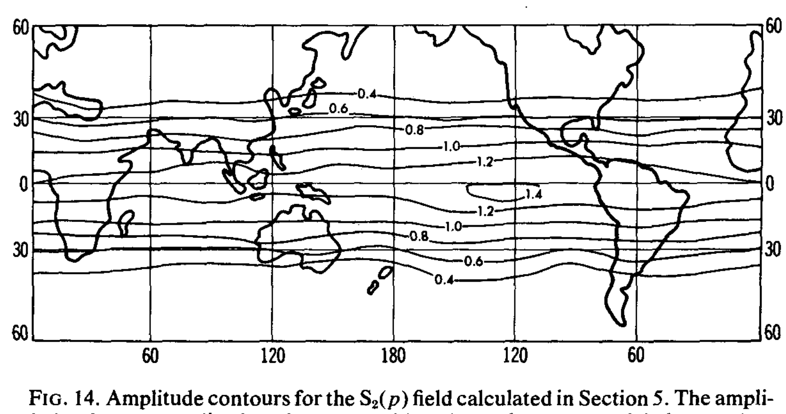

FIG. 14. Amplitude contours for the $S_2(p)$ field calculated in Section 5. The amplitude has been normalized to give reasonable values when contour labels are taken as referring to millibars.

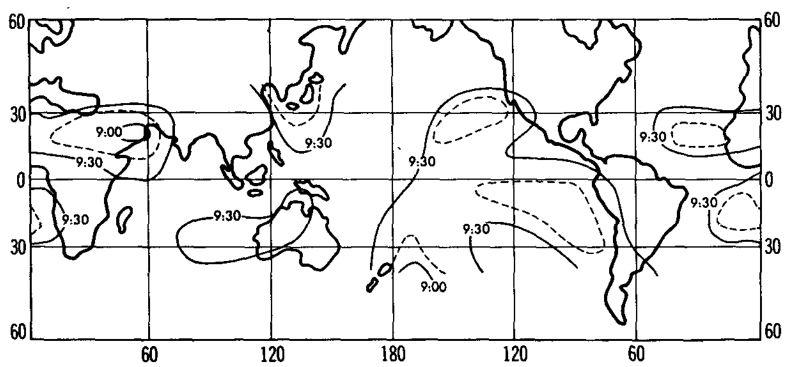

FIG. 15. The 0900, 0915 (dashed) and 0930 contours of the phase of $S_2(p)$ as calculated in Section 5.

Khartoum, Massana and Joal have only very short (less than one year) or unspecified record lengths. In each of these cases, the phase has been determined to be close to 0945 LST. The only station in this dry zone with a determination of $S_2(p)$ based on at least one year's record is Aden (12.8°N, 45.0°E, 1.5 years). Here the phase is anomalously early (about 0924). The stations on the outskirts of this dry region include Tripoli (32.7°N, 13.2°E, 5 years) and Haifa-Jerusalem (32°N, 35°E, 5 years) where the phases are anomalously *late* (1018 and 1022, respectively). In the South Atlantic dry zone, there are two stations listed by Haurwitz, St. Helena (16°S, 5.7°W, unspecified record length) and Ascension (7.9°S, 14.4°W, 2 years). At both stations the observed phase is close to 0945. Haurwitz and Cowley (1968) give some determinations for stations in Australia, but these show no particular anomaly in the phase.[5]

Thus, what data are available generally do not support the existence of anomalously early phases for $S_2(p)$ in the dry regions of the tropics. However, it is not clear that sufficient $S_2(p)$ data really exist to adequately check Lindzen's hypothesis in the manner proposed in this section. There is also a potentially complicating factor in interpreting $S_2(p)$ data over the dry land areas, in that semidiurnal variations in heating due to sensible heat fluxes may be quite large in desert regions and this heating could conceivably mimic the latent heat forcing. It would be most interesting if more determinations of $S_2(p)$ could be made in the dry areas over the oceans.

6. Latent heat release and the 24 h tide

The diurnal surface pressure oscillation, in contrast to $S_2(p)$ is observed to have very large geographical variations. In particular, the amplitude of $S_1(p)$ is observed to be considerably larger at continental stations than at small islands in the open ocean (e.g., Haurwitz, 1965; Haurwitz and Cowley, 1973). The diurnal horizontal wind field is also observed to have large geographical variations which seem to be associated to some extent with the land-sea contrast (Wallace and Hartranft, 1969; Wallace and Tadd, 1974; Yoshida and Hirota, 1979). In addition, the observed magnitudes of $S_1(\mathbf{v})$ in the troposphere (e.g., Hering and Borden, 1962; Hastenrath, 1972; Nitta and Esbensen, 1974) are typically of order 1 m s^{-1}, while classical tidal theory with only zonally-symmetric ozone and water vapor excitations predicts diurnal tropospheric winds which are an order of magnitude smaller (e.g., Lindzen, 1967). The only attempt at explaining these features in the observed diurnal tidal fields appears to be that of McKenzie (1968). In his work McKenzie proposed a simple model for tidal forcing by diurnal variations in turbulent transport of sensible heat within the planetary boundary layer over land surfaces. This model was then employed together with a very simplified land-sea distribution (the boundary-layer forcing over the oceans was taken to be zero) in a three-dimensional classical tidal theory calculation. McKenzie's results were rather discouraging in that he found that diurnal tropospheric winds of the proper magnitude could be produced in his model only by using a boundary layer forcing so strong that the calculated magnitude of $S_1(p)$ was at least twice as large as the observed values.

It is of interest to inquire whether tidal forcing by latent heat release can help to account for the observed features of $S_1(p)$ and $S_1(\mathbf{v})$. In Section 2 it was hypothesized that the diurnal latent heat forcing was much more important (both in terms of amplitude and in terms of phase coherence between stations) over land areas than over the ocean. With this in mind a simple diurnal latent heat forcing model was constructed by setting

$$J(\theta,\phi,z,t) = Af(z)\cos(\theta)\exp[i\sigma(t_{loc} - 18)]$$

[5] It is interesting to note that many of the features of the computed $S_2(p)$ field outside of the dry regions are quite realistic, such as the anomalously early phases over Japan and New Zealand and along the western coast of South America (see Haurwitz, 1956).

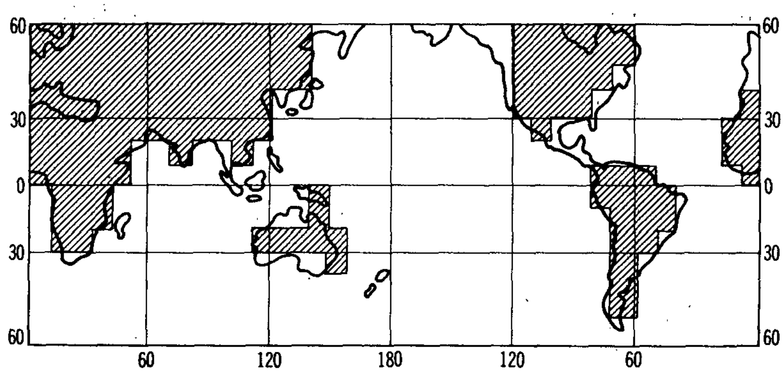

FIG. 16. The simplified land-sea distribution adopted in the diurnal tidal calculations.

over land, and $J = 0$ over oceans, where σ is the angular frequency of the solar diurnal tide, $f(z)$ is taken from Fig. 8 and A is an amplitude that was adjusted so that the magnitude of the forcing at the equator (over land) was the equivalent of a rainfall oscillation of 2 mm day^{-1}. The simplified land-sea distribution adopted is shown in Fig. 16. This was then used in a three-dimensional tidal calculation similar to the one presented in the last Section.[6]

The amplitude of the calculated $S_1(p)$ field is shown in Fig. 17. The pattern is rather complicated but generally the largest amplitudes are seen over the land areas. The phase of the calculated $S_1(p)$ field (not shown) has very large geographical variations (there are several amphidromes) but over the continents the phase tends to be around 0600 LST. This means that the response to latent heat forcing should add in phase with that of the response to O_3 and H_2O forcing (e.g., Lindzen, 1967). This will help to account for the large values of $S_1(p)$ which are observed at continental stations. However, comparison of Fig. 17 with the observations (e.g., Fig. 1 of Haurwitz and Cowley, 1973) shows that the latent heat effect is only one-third to one-half what is actually required to produce the observed geographical variations in $S_1(p)$.

The profiles of the diurnal wind field resulting from the latent heat forcing were calculated for several points. In general, the amplitudes of the calculated $S_1(\mathbf{v})$ field in the troposphere were found to be <0.2 m s^{-1} (i.e., much less than the observed amplitudes). A typical profile is shown in Fig. 18 where it is compared with observations from Washington, DC [the observed profile is taken from McKenzie (1968) and is based on the measurements of Wallace and Hartranft (1969)].

7. Conclusion

Lindzen's hypothesis concerning latent heat release as a possible forcing mechanism for atmospheric tides has been critically examined. In this study it has been necessary to use incomplete data and simplified model calculations. Thus no final verdict can be reached on whether the discrepancies between theory and observations of tidal fields are satisfactorily explained by Lind-

[6] The only difference was that the 740 Hough functions were the ten gravest ones with positive equivalent depths and the 10 gravest ones with negative equivalent depths for wavenumbers −17, −16, ··· 0, 1, ··· 19.

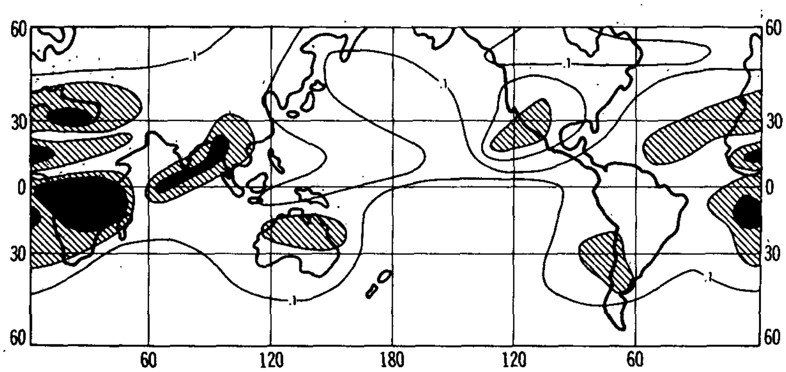

FIG. 17. Amplitude of $S_1(p)$ calculated with latent heat tidal forcing over the continents. Heavy shading indicates amplitudes greater than 0.3 mb, lighter shading indicates amplitudes between 0.2 and 0.3 mb.

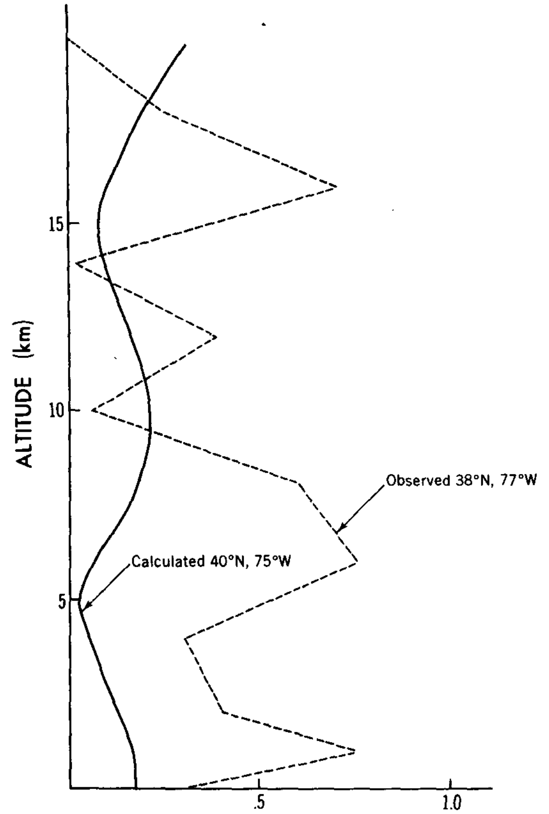

FIG. 18. The calculated amplitude of the meridional component of the diurnal wind oscillation as a function of height at 40°N, 75°W. Shown for comparison are observations at Washington, DC.

zen's hypothesis. The rainfall results of Section 2, however, show that latent heat release is almost certainly an important source of excitation for both the diurnal and semidiurnal atmospheric tides and the observations of $S_2(R)$ in the tropics provide strong support for even the details of Lindzen's theory. The classical tidal theory calculations in this paper have also produced some results which favor Lindzen's ideas. In particular it was shown that 1) latent heat tidal forcing may allow a plausible explanation for the seasonal variation of $S_2(p)$, and 2) the magnitude of the non-migrating components of $S_2(p)$ produced by latent heat tidal forcing is not likely to be unrealistically large. However, when the details of the geographical distribution of the $S_2(p)$ field resulting from latent heat excitation were considered some problems arose. It was shown that the phase of the observed $S_2(R)$ in midlatitudes would lead to a calculated phase of $S_2(p)$ close to 0900 (i.e., earlier than observed). It was also shown that over very dry areas one expects early phases of $S_2(p)$ but that examination of the available $S_2(p)$ station data generally did not reveal anomalously early phases in the dry areas. It was pointed out, however, that neither the $S_2(R)$ nor the $S_2(p)$ data are of sufficiently high quality to regard these discrepancies as fatal for the theory. Finally, it was shown that the diurnal variation of latent heat release is likely to have a significant effect on the 24 h atmospheric tide.

It is not clear how one could improve on the simple theoretical tidal calculations presented in this paper without a better knowledge of the details of the latent heat forcing than is currently available. Thus further progress in studying the role of latent heat in exciting atmospheric tides will likely depend on more analysis of rainfall data. A promising possibility is the use of satellite measurements in such analysis. It is interesting to note that at least two satellite investigations over the tropical Atlantic have found that the diurnal oscillation of IR radiance was much larger than the semidiurnal oscillation (McGarry and Reed, 1978; Murakami, 1979). Finally, it might well be worthwhile to investigate the physics of the diurnal latent heat excitation itself and its interaction with the large-scale dynamics. It is conceivable that there may be other aspects of diurnal variations in the atmosphere (such as the diurnal change in the static stability; e.g., Orlanski, 1973) which must be included in theoretical treatments of latent heat excitation of tidal oscillations.

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APPENDIX A

Reliability of Rainfall Results

Since the rainfall measured at a single station is a quantity with very great variability in time one would expect that very long records of hourly data are needed in order to accurately determine the diurnal or semidiurnal rainfall oscillations. If averages of hourly rainfall rates for individual years were available at a particular station then the determinations of $S_1(R)$ and $S_2(R)$ for each year could be plotted on a harmonic dial and the scatter used to give an indication of the reliability of the mean of all the years (a similar technique is frequently used for assigning error bounds to determinations of $S_1(p)$ and $S_2(p)$; see Chapman and Lindzen, 1970). Unfortunately, the data used in the present study were available only as averages over several years. In a few cases, however, data was available for two or three different periods at the same station (or at stations very close together). The results for these stations are displayed in Table

TABLE A1. Comparison of the observed $S_1(R)$ and $S_2(R)$ at the same (or nearly the same) location for different periods. P is the observed total precipitation rate (mm day^{-1}). A_1 [A_2] is the amplitude (mm day^{-1}) of $S_1(R)$ [$S_2(R)$]. t_1 [t_2] is the phase (local hours) of $S_1(R)$ [$S_2(R)$].

Station	Reference	Latitude	Longitude	Record	P	A_1	t_1	A_2	t_2
Capetown	Schumann	33°54′S	18°32′E	11	1.46	0.35	3.9	0.10	8.6
Capetown	Hastenrath	33°58′S	18°26′E	9	1.36	0.28	5.0	0.08	9.0
Durban	Schumann	29°50′S	31°02′E	13	2.52	1.68	20.9	0.79	8.1
Durban	Hastenrath	29°58′S	30°57′E	10	2.72	1.38	21.4	0.62	8.8
Mombasa	Thompson	4°04′S	39°42′E	5	3.29	2.16	8.4	0.55	0.9
Mombasa	Thompson	4°02′S	39°37′E	16	2.84	1.63	8.9	0.23	0.0
Nairobi	Thompson	1°19′S	36°55′E	6	2.31	1.12	20.6	0.82	4.8
Nairobi	Thompson	1°18′S	36°45′E	9	2.92	1.98	21.0	0.55	5.3
Nairobi	Thompson	1°19′S	36°49′E	2	2.37	1.38	21.7	0.63	5.6
Dar Es Salam	Thompson	6°50′S	39°21′E	9	3.04	2.54	9.7	0.86	11.4
Dar Es Salam	Thompson	6°53′S	39°12′E	9	3.00	2.32	11.1	1.12	0.5
Chukwani	Thompson	6°15′S	39°13′E	11	3.88	2.98	8.9	0.60	11.6
Chukwani	Thompson	6°13′S	39°13′E	12	4.27	3.45	9.2	0.66	11.1

A1. The two different determinations for both Capetown and Durban were made from data taken in non-overlapping periods while for the other stations there was some overlap.

From the table it can be seen that there are quite large (~30%) discrepancies in both the phase and amplitude of $S_1(R)$ and $S_2(R)$ and even in the total precipitation. Thus it seems that to obtain, say, 10% accuracy in $S(R)$ would require records considerably longer than ten years. However, there does seem to be enough consistency between the different determinations of $S(R)$ in the table for one to hope that the analysis of Section 2 (which employs large numbers of stations) may give sensible results.

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