Diurnal Non-Migrating Tides Excited by a Differential Heating Due to Land-Sea Distribution

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Abstract

We have presented a numerical model of non-migrating tides assuming a differential heating localized only on land caused by upward heat flux in the planetary boundary layer. The amplitude of the diurnal temperature oscillation corresponding to the heat source is assumed to be 3 K at the ground at the equator, and to decay exponentially with a scale height of 1 km. Horizontal and vertical structures of non-migrating tides are solved by using classical tidal theory, in which a land-sea distribution is expanded into a complex Fourier series at each latitude with longitudinal wavenumbers ranging from -20 to 20, and further expressed by about 40 Hough modes for each zonal wavenumber. We have found that the horizontal distribution of the heat source has more significant effects on the behavior of non-migrating tides than the vertical distribution.

The model pressure variations of non-migrating tides correlate well with the land-sea distribution at the ground, and agree excellently with observations. They are confined to a region equatorward of 30° latitude, while they spread longitudinally with increasing altitude, which is attributed to the imbalance between the westward and eastward propagating components. Horizontal wind velocities of non-migrating tides are about 2 to 3 m/s near 20 km altitude, and have vertical wavelengths as short as 2 to 5 km, which is fairly consistent with ground-based radar observations.

1. Introduction

Solar¹ atmospheric tides are mainly excited by a heating due to absorption of solar radiation by water vapor in the troposphere as well as ozone in the middle atmosphere, and eddy heat transfer near the ground. When these heat sources are uniformly distributed with longitude, atmospheric tides are zonally homogeneous and propagate westward with the sun, and are called migrating tides. Numerous observations and theoretical studies on the migrating tide, especially diurnal and semidiurnal components, have been continued over many decades [Chapman and Lindzen, 1970; Forbes and Garrett, 1979; Kato, 1980].

Non-migrating tides excited by localized heat sources also have wave periods corresponding to a subharmonic of a solar day, but they have various longitudinal wavenumbers, and therefore, they do not necessarily propagate westward with the sun.

Observationally, the existence of non-migrating tides has been known for more than half century as reviewed by Chapman and Lindzen [1970]. A zonally standing oscillation is a main component of semidiurnal surface pressure variation in the polar

region [Simpson, 1918; Haurwitz, 1956], although the migrating tide becomes dominant below 50° latitude. Haurwitz [1965] detected non-migrating tides in diurnal surface pressure oscillation, which have larger amplitudes on land than over ocean. From radiosonde observations, Wallace and Tadd [1974] found that diurnal wind oscillations are largely affected by topography, and are characterized by a vertical wavelength shorter than 10 km.

Recent MST radar observations have revealed characteristics of quasi-inertia gravity waves. Some of these waves are interpreted as diurnal tides because of their repeatability fixed with local time. Fukao *et al.* [1978] detected a diurnal oscillation in the lower stratosphere by the Jicamarca VHF radar with a height resolution of 4 km, which has horizontal amplitudes ranging from 1 to 5 m/s with a maximum occurring near 20 km. Although a vertical wavelength was difficult to determine because of the poor height resolution, it was estimated as shorter than 10 km. Arecibo UHF radar observations have detected a similar diurnal oscillation at 10 to 30 km, which has a horizontal amplitude of about 2.5 m/s with a vertical wavelength of 3 to 5 km [Fukao *et al.*, 1980]. Tsuda *et al.* [1985] also observed vertical profiles of a wave with amplitude

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and vertical wavelength of 2 to 4 m/s and 3 km, respectively. This wave was detected on 2 successive days synchronizing to the local time, so that it was interpreted as tides.

The behavior of non-migrating tides have been numerically studied by using the classical tidal theory for various heat source models. McKenzie [1968] investigated non-migrating tides assuming an empirical heat source, and successfully explained fundamental features of observations: large wind amplitudes below 30 km and considerable attenuation of non-migrating tides with latitude, although their fine structures were not fully delineated mainly due to severe truncation in Hough mode expansion and simplified latitudinal and longitudinal structures of the heat source model.

Giwa and Hussain [1978] developed a numerical model of a semidiurnal standing oscillation in the polar region assuming the eddy heat transfer localized on land.

Kato *et al.* [1982] studied thermal excitation of diurnal non-migrating tides by assuming idealized heat sources localized near the equator, and found that non-migrating tides generally have short vertical wavelengths consistent with the Arecibo and Jicamarca observations [Fukao *et al.*, 1978, 1980].

Using an equivalent gravity wave f-plane model Forbes and Groves [1987] have calculated the structure of diurnal propagating tides at low latitudes including the excitation of non-migrating tides due to longitudinal variations in insolation absorption by tropospheric water vapor. They have found that the non-migrating tides can give longitudinal variations of at least ± 12 to 15 % about zonal mean values in the diurnal wind amplitude at low latitudes at 80 to 100 km.

Migrating tides are detected in a general circulation model (GCM) which is fundamentally employed to simulate large scale dynamics of the earth's atmosphere [Hunt and Manabe, 1968]. Recently, the non-migrating tides are also studied by using sophisticated GCM with improved heat sources [Zwiers and Hamilton, 1986; Tokioka and Yagai, 1984, 1987]. Especially, Tokioka and Yagai [1987] have extracted the detailed structure of heat sources of atmospheric tides from the GCM that includes realistic surface boundary conditions such as the land-sea distribution, topography, snow coverage and sea-ice distribution, and found that the dominant heat source of the diurnal non-migrating tides is the eddy heat transfer within the planetary boundary layer. However, the vertical resolution of the present GCM might not be good enough in expressing the structure of non-migrating tides with vertical wavelengths shorter than 10 km [Tokioka and Yagai, 1987].

This paper is concerned with a numerical model of diurnal non-migrating tides excited by upward heat

flux within the planetary boundary layer localized only on land, which has been found as the major excitation source of the diurnal non-migrating tides by Tokioka and Yagai [1987]. We have used the classical tidal theory including Newtonian cooling in order to express the non-migrating propagating modes that are characterized by vertical wavelengths as short as down to a few kilometers, because this method can obtain a fairly good height resolution (40 m in the present case [Kato *et al.*, 1982]). The classical tidal theory, however, is defective in including a variety of excitation sources with realistic vertical and horizontal distributions.

An extended abstract of this study has been reported by Tsuda and Kato [1985].

2. Heat source model

The heat source can generally be expressed by multiplying the intensity of incident solar flux and amount of the absorber. In order to simplify the numerical computation in the present work we treat separately the vertical and horizontal distributions of the heat source. A 24-hour component of the heat input per unit mass of the atmosphere in unit time is expressed as follows [Kato *et al.*, 1982]:

$$J = aS(\theta, \phi)Z(z) \sin \theta \exp(i(\Omega t + \phi)) \quad (1)$$

where θ, ϕ, z, t and Ω , are the co-latitude, longitude, altitude, universal time and Earth's angular velocity (c/day), respectively. $S(\theta, \phi)$ and $Z(z)$ represent horizontal and vertical distributions of the heat source. a is a constant which determines the maximum heat input at the ground at the equator. The fourth and fifth factors in (1) correspond to the latitudinal decrease of the incident solar flux under equinoctial condition, and diurnal variation in terms of the local time, respectively. Now, we need to define $S(\theta, \phi)$ and $Z(z)$ in order to solve the problem numerically.

2.1 Vertical structure

First, we investigate a vertical profile of the upward heat flux over land whose vertical derivative divided by the atmospheric density corresponds to the heat input. Caughey and Palmer [1979] determined a height structure of normalized heat flux near the ground, which suggests that the decrease of the heat flux can be approximated by $\exp(-z/(0.5z_i))$ or a little more gradual, where z_i is a maximum height of the mixing layer.

Carson [1973] determined z_i as high as 1800 m, and also deduced diurnal variation of the heat flux at ground, which showed a sinusoidal variation during daylight hours with a maximum of about 40 mW/cm² at noon, and small negative values at night. The maximum heat flux seems to become

Fig. 1. Latitudinal distributions of Fourier expansion coefficients of $S(\theta, \phi) \sin \theta$ for $|\tilde{m}| = 0$ to 4.

somewhat larger at the equator because of more intense incident solar flux. We assume in this paper that the maximum heat flux at the equator is larger than the Carson's estimation by about 30 %, which implies a maximum heat input of 433 mW/kg at the ground or amplitude of a 24-hour component of the corresponding temperature oscillation of 3 K.

On the other hand, the diurnal variation of air temperature above the sea is primarily controlled by the diurnal variation of the sea-surface temperature, and has amplitudes ranging from 0.1 to 0.8 K depending on the activity of convection and turbulence [e.g., Roll, 1965]. By comparing the amplitudes of the diurnal temperature variations between on land and at sea, we can assume that a significant differential heating is localized only on land, *i.e.*, $S(\theta, \phi) = 1$ and 0 on land and ocean, respectively.

We also assume that the heat input has a sinusoidal variation with a maximum occurring at 12 LT, and decreases exponentially at a scale height of 1 km from the surface value without change in the maximum phase. This heat source model is mainly used in the later sections, although another model with a scale height of 0.5 km is additionally used for a comparison.

2.2 Fourier and Hough mode expansion

The horizontal distribution of the heat source $S(\theta, \phi) \sin \theta$ with a resolution of 1° for both latitude and longitude is expanded into complex Fourier series for longitudinal wavenumbers ($\tilde{m}$) ranging from -21 to 19. Since the heat source distribution is a real function, the coefficients for negative $\tilde{m}$ become a complex conjugate to those for positive $\tilde{m}$. Fig. 1 shows latitudinal distribution of amplitudes of Fourier coefficients for $|\tilde{m}| = 0$ to 4.

Note that wavenumbers of the Fourier components for diurnal tides are shifted by one from those for the land-sea distribution, because (1) includes the factor of $\exp(i\phi)$ attributed to local time dependency of the solar incident flux [e.g., Kato, 1980]. Therefore, for example, the component with $\tilde{m} = 0$ excites a diurnal tide propagating westward with a longitudinal wavenumber of 1, that is, the migrating tide, and that with $|\tilde{m}| = 4$ corresponds to westward and eastward propagating diurnal tides with wavenumbers of 5 and -3, respectively [Tokioaka and Yagai, 1987]. Now, we define the longitudinal wavenumber for diurnal tides as $m = \tilde{m} + 1$.

We investigate here the distribution of Fourier coefficients shown in Fig. 1. The coefficient for $\tilde{m} = 0$ ($m = 1$) is closely related to the distribution of continents and is large at $20^\circ - 70^\circ\text{N}$, shows considerable value at $20^\circ\text{N} - 30^\circ\text{S}$, and becomes fairly small at $35^\circ - 65^\circ\text{S}$. Although we assume the longitudinally inhomogeneous heat source for the excitation of non-migrating tides, this Fourier component exceptionally generates the migrating tide.

We can recognize further that coefficients for $|\tilde{m}| = 1$ to 4 are generally large at $60^\circ\text{N} - 30^\circ\text{S}$, and become very small at $35^\circ - 65^\circ\text{S}$, which is consistent with the behavior of the thermal forcing determined from GCM [Tokioaka and Yagai, 1987]. Therefore, the non-migrating tides are more excited in the former region, while in the latter region the migrating tide is mainly generated by a homogeneous heat source. The coefficients for $|\tilde{m}| = 1, 2$ and 4 are specifically large at $5^\circ - 30^\circ\text{N}$, $30^\circ - 60^\circ\text{N}$, and $5^\circ\text{N} - 30^\circ\text{S}$, respectively, while the component for $|\tilde{m}| = 3$ has broad enhancement at $60^\circ\text{N} - 30^\circ\text{S}$.

The latitudinal distributions of the complex Fourier coefficient for $m = -20$ to 20 are further

Fig. 2. Eigenvalues of the Hough mode versus longitudinal wavenumber m . Full and open circles correspond to symmetric and antisymmetric modes, respectively.

expanded into a series of Hough modes. Usually 40 Hough functions corresponding to 20 positive and 20 negative equivalent depths are used in the latitudinal expansion except for $m = 0$ and 1 which include additionally a mode with an infinite equivalent depth.

The distribution of eigenvalues for $m = -10$ to 10 is plotted in Fig. 2, where each branch is labeled according to the mode number of the migrating diurnal tide [Kato, 1980]. Except for $n = -1$ and -2 , symmetric and antisymmetric negative modes have eigenvalues very close to each other, while the positive modes have quite different eigenvalues between the symmetric and antisymmetric modes. Furthermore, the distributions of positive and negative eigenvalues are roughly symmetric relative to the values for $m = 1$ and 0, respectively. However, precise investigation shows discrepancies from the symmetry such that the positive eigenvalues for $m > 1$ are slightly smaller than those for $m < 1$, and negative ones are smaller for $m < 0$ than $m > 0$. Since a vertical wavelength of the propagating (positive) mode becomes shorter as the eigenvalue becomes larger [e.g., Kato 1980], the eastward propagating mode ($m < 0$) generally has smaller vertical scale than the corresponding westward propagating wave ($m > 0$). Note that branches for $n = -1$ and -2 plotted in Fig. 2 have positive eigenvalues for $m \leq 0$ and $m \geq -1$, respectively, which produces large asymmetry in the distribution of the positive eigenvalues relative to those for $m = 1$.

By using Hough functions corresponding to the eigenvalues plotted in Fig. 2, the heat input J can

finally be expressed as follows:

$$J = a \sum_{m=-20}^{20} \sum_{n=-20}^{20} C_m^n \Theta_m^n(\theta) \exp(im\phi) \exp(-z/H_s) \exp(i\Omega t) \quad (2)$$

(n ranges from -21 to 20 for $m=0$ and 1)

Fig. 3 shows a recomposed horizontal distribution of J at the ground at $t=0$, which approximates fairly well the land-sea distribution. The horizontal distribution of the heat source consequently reflects a coastal effect as a result of truncation of the Fourier series, i.e., the heat source intensity is large in the middle of continents, and is reduced near the coast. If the number of Fourier components is reduced, details of the land-sea distribution disappear. We have found that only major continents are outlined using $\tilde{m} = -5$ to 5 . The continents are fairly well approximated for $\tilde{m} = -10$ to 10 , although the coastal effect can be recognized even in the center of the continents. By increasing the range of $\tilde{m}$ to -20 to 20 , we can reproduce relatively fine details of the topography including some peninsulas and islands.

Vertical structures of more than 1600 tidal modes are solved using a similar numerical technique described by Kato *et al.* [1982] to which a radiative damping process is added assuming a Newtonian cooling model proposed by Blake and Lindzen [1973], although this process is not significant below 30 km.

It is noteworthy that we do not include zonally homogeneous heat sources due to water vapor and ozone, which are important in generating the migrating tide, because we are devoted to extracting

Fig. 3. Horizontal distribution of a recomposed heat source.

the excitation of non-migrating tides by the localized heat source. However, since the classical tidal theory used in this paper assumes linear equations, a total tidal field could be estimated by combining models of migrating and non-migrating tides.

3. Results and discussions

3.1 Horizontal distribution of perturbed pressure

Amplitudes of the perturbed diurnal pressure oscillation p_1 are plotted in Fig. 4. Contour lines are drawn at every one-fifth of the maximum p_1 at each altitude, which are 120, 30 and 4.3 N/m² at ground, 11 and 30.3 km, respectively. The decrease of these maxima in p_1 is approximately exponential with a scale height of about 9.4 km. Note that when the migrating components of p_1 were plotted in Fig. 4, they become straight lines parallel to the equator, because they do not have any longitudinal variation, although there is a latitudinal distribution.

At the ground, p_1 is clearly affected by the land-sea distribution, and its enhancement is conspicuous over the equatorial regions of the continents, especially over Africa and South America. A comparison of this distribution with observations by Haurwitz [1965] is discussed later in this subsection.

At 11 km p_1 consists of many small patchy structures elongated with longitude, which are converged latitudinally into a fairly narrow region near the equator. Although each wave component with different zonal wavenumber propagates either westward or eastward, non-migrating tides as a whole appear to be like localized standing oscillations, which are restricted at longitudes corresponding to the region of large p_1 at the ground.

Direct influence of the localized heat source is negligible at 11 km. Therefore the negative (evanes-

cent) modes are not excited at this altitude, but the positive modes propagating from the lower atmosphere become dominant. The propagating modes are fundamentally confined in a latitudinal region lower than 30° where the inertial period is longer than the wave period of the diurnal tide, and have an oscillatory latitudinal distribution with a number of nodes equal to $n+1$ [e.g., Kato, 1980]. Characteristics of p_1 at 11 km are consistent with the behavior of the higher order modes of the propagating tides.

At 30.3 km patchy patterns of p_1 are also confined to be near the equator, but are more uniformly diffused with longitude than those at 11 km. There seems to be a general tendency that p_1 spreads longitudinally as it separates from the heat source region, which can be explained in terms of the difference in structures of the positive modes. If a pair of westward and eastward propagating tides with the same absolute zonal wavenumber are excited by the same heat source intensity, they will interfere and cancel each other to produce a zonally standing oscillation. However, the distributions of the excitation source for the pair of waves are quite different, because the heat sources of these two waves correspond to different $\tilde{m}$ as described in the previous section. Moreover, their eigenvalues are slightly different as shown in Fig. 2, which gives difference in the vertical structures as well as the latitudinal distributions. Because of these imbalances between the behavior of the westward and eastward propagating modes, which become larger as altitude becomes higher, p_1 tends to propagate longitudinally in the direction of the component with larger amplitude between the pair of the waves.

Haurwitz [1965] observed a global distribution of the diurnal surface pressure oscillation, and sepa-

Fig. 4. Contour plots of amplitudes of the perturbed pressure at ground (bottom), 11 km (center) and 30.3 km (top). Contour lines are drawn at intervals of one-fifth of the maximum value.

Fig. 5. Amplitudes of the sine component of the diurnal surface pressure oscillation reproduced from observations by Haurwitz [1965] (top) and numerical results in the present study (bottom). Contour lines are plotted at intervals of 0.2 mmHg.

Fig. 6. Fourier coefficients of the diurnal surface pressure oscillation averaged over entire latitudes for observations (top; reproduced from Haurwitz [1965]), and the present study (bottom). Coefficients are obtained for $m = -7$ to 9 and $m = -20$ to 20 for observations and the numerical model, respectively.

rated it into sine and cosine components as a function of local time. (Note that these components are mislabeled in Haurwitz's paper, as pointed out by Chapman and Lindzen [1970].) Fig. 5 shows a comparison of the sine component of the surface pressure variation p_1^s between the observations and the numerical model, where contour lines are drawn every 0.2 mmHg (about 26.7 N/m²) adjusting to the results by Haurwitz.

The observed p_1^s is generally positive, and becomes as large as 0.8 mmHg over Africa, South America and South-Eastern Asia, and is also considerably enhanced over Australia. These characteristics are quantitatively reproduced by the numerical model except that the p_1^s around South-Eastern Asia is slightly smaller in the present computation.

As stated in Section 2, we do not include the excitation source distributed uniformly with longitude. Therefore, the model includes another discrepancy in the behavior of the migrating component, which is particularly recognizable as contour lines labeled as 20 at 40° – 60° N and 40° S– 60° S in Fig. 5. If we estimate the amplitude of the migrating component

as 0.2 mmHg, the heat input assumed in the present study might be slightly overestimated by 20 – 30 %.

Fig. 6 shows Fourier coefficients of p_1^s integrated over latitudes determined from the observations [Haurwitz, 1965] and the model. The observations show large coefficients ranging from 5 to 10 N/m² for $m = 3, -1, 2, 3$ and 5 . The coefficient for $m = 1$ is 38 N/m², which exceeds the model by about 30 N/m² because of the reason described above. The coefficients for the model at $m = -3$ to 0 and 2 to 5 have similar amplitudes to the observation, although correlation for each wavenumber is not perfect.

3.2 Wind velocity profile

Considering that above the heat source region non-migrating tides have significant amplitudes only equatorward of 30° latitude, we describe wind velocity profiles at 20° N and 70° W determined by the present model, which can be compared with the UHF radar observation at Arecibo (18° N, 67° W). Fig. 7 shows vertical profiles of amplitude and phase of the eastward wind velocity deduced from the model together with the observation collected at

Fig. 7. Vertical profiles of amplitudes (left) and phases (right) of the eastward component of diurnal tidal winds. The phase is defined as local time of the eastward maximum wind velocity. A solid line and a chained line with an open circle correspond to the numerical results of the present study at 70°N and 20°N and a theoretical model of migrating tides at 18°N (after Forbes [1982]), respectively. A dashed line with dots shows radar observations at Arecibo (18°N , 67°W) (reproduced from Fukao *et al.* [1980]).

Arecibo [Fukao *et al.*, 1980] and a theoretical model of the migrating tide at 18°N [Forbes, 1982]. Note that Fig. 7 does not necessarily show a typical profile at 20°N , because the non-migrating tides have both latitudinal and longitudinal variations, while the migrating tide represented by the Forbes' model is homogeneous with longitude.

Our model shows that the vertical wavelength is about 5 km below 15 km altitude, becomes as short as 2 to 3 km at 15 to 30 km altitudes, then finally becomes longer than about 10 km above 30 km. Fluctuations in the amplitude profile correspond well to the phase structure. The maximum amplitude is 1.1 m/s and 3.5 m/s in the regions below 15 km and at 15 to 30 km, respectively, reaches 8 m/s at 35 km, and becomes attenuated above 35 km.

These characteristics suggest that the propagating modes with relatively long vertical wavelength (5 km) are dominant below 15 km, while higher order modes with shorter vertical wavelengths (2 to 3 km) grow as altitude increases, and are dominant above 15 km. A complicated phase profile at 15 to 20 km altitudes seems to exhibit an interference between waves with different vertical wavelengths. Because the dissipative process becomes effective above 30 km, the modes with the short vertical wavelengths are removed, and the vertical wavelength becomes longer with altitude.

The model for the migrating tide predicts a fairly constant vertical wavelength of 25 km over the entire height region, which suggests a manifestation of the fundamental symmetric mode ($S_{1,1}$ mode). The amplitude generally increases from a few tenths of meters per second near the ground to about 40 m/s at 84 km, and is smaller than the non-migrating tides below about 42 km. Discrepancies between the two models below 30 km altitude are characterized such that the non-migrating tides have shorter vertical wavelength (1/5 to 1/10) and larger amplitudes (up to 10 times) than those for the migrating tide.

Fig. 7 also shows amplitudes and phases of the eastward component of diurnal oscillations detected by the Arecibo UHF radar observations [Fukao *et al.*, 1980]. Amplitudes are determined as large as 2.5 to 3 m/s at 17 to 21 km altitude, and become smaller in regions above and below this layer, although their exact values were not delineated. The observed amplitudes are consistent with our model, and considerably larger than the migrating component. Phases are delineated at 13 to 28 km at an interval of 1 km. Our model explains the observed short vertical wavelength fairly well and also shows relatively good agreement with the absolute values of the observed eastward phase.

The southward component of non-migrating tides at 20°N and 70°W is shown in Fig. 8, in which H_s is

assumed to be 0.5 km as well as 1 km. Overall structures of the amplitude and phase profiles for $H_s = 1$ km are quite similar to those for the eastward component except that phases are rotated by 6 hours (90°), which is consistent with the quadrature relation between zonal and meridional components of tidal winds [*e.g.*, Kato 1980].

The phase profile for $H_s = 0.5$ km agrees well with that for $H_s = 1$ km except for fine structures at 30 to 40 km. Although the wind amplitudes at ground are not largely different between the two cases, those for $H_s = 0.5$ km become about half comparing the $H_s = 1$ km case above the heat source region. Since the total heat input is proportional to aH_s , it is suggested that the amplitudes of the non-migrating tides in a region above the heat source are proportional to the integrated heat input.

It is noteworthy that the change in the height scale of the heat source from 1 to 0.5 km is not effective in modifying the vertical structures of non-migrating tides except for their amplitudes. The dominant mode with a specific vertical structure corresponds to a latitudinal structure characterized by a Hough function, whose excitation is substantially controlled by the horizontal distribution of the heat source. As a conclusion, the horizontal distribution of a heat source is more important than its vertical variation in determining the vertical structure of non-migrating tides as well as their horizontal structure.

In Fig. 8, also plotted are amplitudes of non-migrating tides assuming another model by McKenzie [1968] (hereafter referred to as MK68), which used a numerical technique similar to the present study. It might be useful to review briefly the heat source model used by McKenzie, in which the horizontal distribution is expressed by a multiplication between independent longitudinal and latitudinal distributions shown in Fig. 9. MK68 also assumed that a temperature oscillation corresponding to the heat input exponentially decreases at a scale height of 1.25 km from 8 K at the ground.

Now, we compare MK68 with our model for $H_s = 1$ km. Since the amplitude of the heat source at ground is $8/3$ times larger for MK68, the wind amplitude at ground shown in Fig. 8 is larger than our model. However, in the height region from just above the heat source up to about 45 km, MK68 shows considerably smaller amplitudes than our model, although the total heat input is $10/3$ times larger in MK68. It is suggested that the simple horizontal distribution of the heat source shown in Fig. 9, especially the broad latitudinal variation, has mainly generated vertically evanescent modes which stay within the heat source region, and does not excite significant propagating modes. This comparison also confirms that the horizontal distribution of the heat source is essential in determining the behavior

of the non-migrating tides.

4. Concluding remarks

This paper is concerned with a numerical model of diurnal non-migrating tides using the classical tidal theory, in which the heat source is assumed to be localized only in the planetary boundary layer on land. The main conclusions are summarized as follows:

- (1) At the ground, p_1 is clearly affected by the land-sea distribution with large amplitudes over the equatorial regions of the continents, which agrees well with the observations by Haurwitz [1965].
- (2) Fourier coefficients of p_1^s have large amplitudes for m ranging from -3 to 5 .
- (3) Above the heat source region, p_1 has complicated patchy patterns elongated in the longitude, and is confined equatorward of 30° latitude, which can be interpreted in terms of the characteristics of Hough modes.
- (4) Wind velocity profiles of the non-migrating components are characterized by short vertical wavelengths (2 to 5 km) and fairly large amplitudes (3 m/s near 20 km) in comparison with the migrating component, which are consistent with radar observations at Arecibo [Fukao *et al.*, 1980; Tsuda *et al.*, 1985].
- (5) The horizontal distribution of the heat source is more important than the vertical distribution in determining the vertical structure of the non-migrating tides as well as their horizontal structure.
- (6) When the horizontal distribution of the heat source is specified, the amplitude of the non-migrating tides above the heat source is determined by the total heat input which is proportional to aH_s .

In this paper, we have extracted a sinusoidal variation of the heat source with the period of 24 hours from the diurnal variation of vertical heat flux in the boundary layer, and studied the excitation of the forced oscillations with discrete zonal wavenumbers and specific latitudinal structures. However, the diurnal variation of the heat flux consists of many sub-harmonics of a solar day. Furthermore, if the vertical structure of the heat flux has day-by-day variation depending on activities of turbulence and convection, it can have various periods longer or shorter than 24 hours. Therefore, the localized heat source in the boundary layer on land does not exclusively

Fig. 8. The same as Fig. 7 except for the southward component. Solid and dotted lines correspond to the model calculations with $H_s=1$ and 0.5 km, respectively. The dashed line shows the amplitudes of non-migrating tides calculated by McKenzie [1968].

Fig. 9. Longitudinal (top) and latitudinal (bottom) distributions of the localized heat source model assumed by McKenzie [1968]. A smoothed line in the top panel is used in the MK68 model.

excite the diurnal non-migrating tide, but could generate many waves with various periods and horizontal wavenumbers. Therefore, it is suggested that the heat source studied in the paper might be responsible, at least in part, for the generation of long-period gravity waves in the troposphere and lower stratosphere. Such waves have been theoretically determined to play an important role in transporting momentum and energy from the lower atmosphere to the middle atmosphere and decelerating general circulations [e.g., Lindzen, 1981; Fritts, 1984].

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海陸分布に起因する差分熱源による太陽に同期しない
一日周期大気潮汐波に関する数値計算の研究

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地球大気の境界層中の顕熱による太陽非同期の一日周期大気潮汐波に関する数値計算を行った。ここで熱源は大陸上にのみ局在するとし、これに伴う温度変動は赤道上の地表付近で 3 K であるとし、高度とともにスケールハイト 1 km で指数関数的に減少するとした。海陸分布は緯度 1°毎に経度波数 $-20 \sim 20$ のフーリエ成分に分解され、さらに 40 項の Hough モードで展開された。これらをもとに古典潮汐論を用いて一日周期潮汐波の鉛直方向および水平構造を求めた。

この結果太陽非同期の大気潮汐波の振舞いは熱源の鉛直分布よりむしろ水平分布により大きな影響を受けることが分かった。地表での一日周期大気潮汐波による大気圧変動は観測とよく一致することが分かった。高度が高くなるにつれてこの変動は南北緯 30°以内に限定されるとともに経度方向に広がってくることが明らかになった。これは東向きおよび西向き伝搬波の高度構造が若干異なるためにこれらの波の間のバランスが保てなくなり優勢な波の方向に伝搬することによると考えられる。太陽非同調の一日周期潮汐波の水平成分の振幅は高度 20 km で 2 ~ 3 m/s であり、鉛直波長は 2 ~ 5 m/s であることがこの計算で示されたが、これらは大型レーダーによる観測結果とよく一致している。