6.1. Introduction

A simulation of diurnal tides corresponding to the latitude of Gadanki is done using classical tidal theory and taking into account major heat sources, which are believed to contribute to the nonmigrating tidal fields along with the water vapour heating, which is predominantly responsible for the migrating tidal components. The heat sources considered for the simulation were mainly solar radiation absorption by water vapour in the troposphere, planetary boundary layer heat flux, and latent heat release in deep connective clouds. Recent observational studies (Cess et al., 1995; Ramanathan et al., 1995; Pielewskie and Valero, 1995) suggest that more enhanced heating of the clouds by solar short-wave radiation takes place in the troposphere than previously thought. This enhanced absorption by clouds amounts to globally averaged value of \(-25\) W m\(^{-2}\). In view of this fact the cloud heating assumes significance for the generation of tidal oscillations (Braswell and Lindzen, 1998). In a numerical study, Braswell and Lindzen (1998) showed that inclusion of anomalous absorption of solar radiation by clouds could remove the long-standing discrepancy of diurnal migrating surface pressure amplitudes predicted by classical tidal theory. It is possible that this excess cloud heating could generate nonmigrating tidal modes also. For the present simulation cloud heating is also included along with the other three sources. The simulated amplitudes and phases are compared with observed amplitudes and phases of diurnal oscillations over Gadanki.

6.2. The Classical Tidal Theory

Chapman and Lindzen (1970) describes the theoretical understanding of the tides, indicating the shortcomings of the theory and the likely directions of future progress. According to them the tides are produced as a result of thermal forcing of
water vapour in the atmosphere by the solar radiation. For the forcings, one has to specify the period, phase, amplitude, vertical distribution and horizontal distribution. In respect of the horizontal distribution, no account is taken of land-sea contrasts or orography. Hence the forcing is assumed to be symmetrical along a latitude circle. Therefore, for horizontal distribution, it is enough to specify the distribution with respect to latitude only.

Chapman and Lindzen specify different forcings for the 24-hour and the 12-hour periods. Vertical distribution of thermal forcing is considered to be the same both for 24-hour and 12-hour oscillations. Latitudinal distribution is considered to be the same for 24-hour as well as 12-hour oscillation, although amplitudes and phases are different. Other assumptions made by Chapman and Lindzen are:

i) The motion of the atmosphere may be described by the Navier-Stokes equations for a compressible gas. It is convenient to express them in spherical co-ordinates for a frame of reference rotating with the earth.

ii) The atmosphere is always in local thermodynamic equilibrium, i.e., it responds to heating via a continuous sequence of equilibrium states.

iii) Gas constant $R$ is the same throughout the atmosphere, i.e., the atmosphere is a perfect gas.

iv) Gravitational acceleration $g$ is constant in the horizontal and along the vertical.

v) The earth is taken as a sphere, without ellipticity and without orography.

vi) The atmosphere is taken to be in hydrostatic equilibrium.

vii) Dissipative process such as molecular viscosity, turbulent eddy viscosity, thermal conductivity, ion drag and infrared radiative transfer are ignored.

viii) Tidal fields are considered as linearisable perturbations on the basic state.

Let a given field $f$ may be written

$$f = f_0 + f'$$

where $f_0$ is the basic field and $f'$ is the tidal contribution. By this 'linearisation' one may neglect quadratic and higher order terms in $f$. If there is a tidal excitation $E$, then $f'$ is proportional to $E$. 

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ix) To make the tidal field equations more tractable, the basic fields are assumed to be steady and the basic flow may be set to zero. Thus the basic temperature, pressure and density are independent of longitude and latitude.

Making use of perfect gas law, hydrostatic approximation, conservation of momentum equations (three components), the first law of thermodynamics for an ideal gas and continuity equation and assuming the tidal fields to be linear and varying harmonically in time and longitude one arrives at a single partial differential equation in spherical coordinates,

\[
\frac{\partial^2 G^{n,s}}{\partial z^2} + \left( \frac{dH}{dz} - 1 \right) \frac{\partial G^{n,s}}{\partial z} = \frac{g}{4a^2 \omega^2} \left[ \left( \frac{dH}{dz} + \kappa \right) G^{n,s} - \frac{\kappa J^{n,s}}{\gamma g H} \right]
\]

where

\[
G = -\frac{1}{\gamma} \frac{DP}{P_0} \frac{DP}{Dt} \tag{6.2}
\]

\[
\frac{DP}{Dt} = \frac{\partial \delta P}{\partial t} + \frac{dP_0}{dz}
\]

\[
= \gamma g H \frac{DP}{Dt} + (\gamma - 1) \rho_0 J \tag{6.3}
\]

\[
P = P_0 + \delta P
\]

\[
\rho = \rho_0 + \delta \rho \tag{6.4}
\]

The notations used

- \(P_0, \rho_0\) basic pressure and density fields
- \(\delta P, \delta \rho\) tidal perturbations in the pressure and density fields
- \(t\) time
- \(\gamma = c_p/c_v = 1.4\), ratio of specific heats
- \(c_v, c_p\) specific heats of air at constant volume and pressure respectively
- \(z\) vertical co-ordinate, distance from the earth's surface in the vertical direction
$u, v, w$ southward, eastward and upward component of tidal perturbation in the velocity field

$H = R T_0 / g$ scale height

$T_0$ basic temperature field

$g$ acceleration due to gravity

$R$ gas constant for air, assumed to be constant for the atmosphere

$a$ radius of the earth

$\omega$ angular frequency of the earth's rotation

$\kappa = \gamma^{-1}/\gamma = 2/7$

$m$ angular frequency of tidal oscillation

$s$ zonal wavenumber (in longitude)

$F$ is an operator and

$$F = \frac{1}{\sin \theta} \frac{\partial}{\partial \theta} \left( \frac{\sin \theta}{f^2 - \cos^2 \theta} \frac{\partial}{\partial \theta} \right) - \frac{1}{f^2 - \cos^2 \theta} \left( s \frac{f^2 + \cos^2 \theta}{f^2 - \cos^2 \theta} + s^2 \frac{1}{\sin^2 \theta} \right)$$

(6.5)

$\theta$ co-latitude

$f = m/2\omega$

$J$ thermodimensional heating per unit mass per unit time

$J$ and $G$ vary in time and longitude as

$$J = J^{m, s} e^{i (n s + \phi)}$$

$$G = G^{m, s} e^{i (n s + \phi)}$$

(6.6)

The second order partial differential equation (6.1) can be solved by the technique of separation of variables. The variables $J$ (forcing) and $G$ (response) are functions of co-latitude $\theta$ and height $z$. The solution is of the form
In equation (6.7) \( L_n^{m,s} \) is the vertically dependent part and \( \Theta_n^{m,s} \) the latitudinally dependent part. Since \( \Theta_n^{m,s} \) are orthogonal functions for each \( m \) and \( s \), \( L_n^{m,s}(z) \Theta_n^{m,s}(\theta) \) will also be a solution of equation (6.1).

Similarly for \( J \), we have

\[
J_n^{m,s} = \sum_n J_n^{m,s}(z) \Theta_n^{m,s}(\theta)
\]  

(6.8)

Making use of equations (6.7) and (6.8), equation (6.1) is now separated into two parts.

\[
F(\Theta_n^{m,s}) = -\frac{4a^2\omega^2}{gh_n^{m,s}} \Theta_n^{m,s}
\]

(6.9)

and

\[
H \frac{d^2 L_n^{m,s}}{dz^2} + \left( \frac{dH}{dz} - 1 \right) \frac{dL_n^{m,s}}{dz} + \frac{1}{h_n^{m,s}} \left( \frac{dH}{dz} + \kappa \right) L_n^{m,s} = \frac{\kappa}{\rho g h_n^{m,s}} J_n^{m,s}
\]

(6.10)

\( h_n^{m,s} \) is the constant of separation.

In equation (6.9) \( \{h_n^{m,s}\} \) are the eigen values and \( \{\Theta_n^{m,s}\} \) are the eigen functions. Equation (6.9) was first derived by Laplace for the free surface oscillations of a spherical ocean envelope and so is called Laplace’s tidal equation. However, in his original formulation \( h_n^{m,s} \) was replaced by the depth of the ocean and \( h_n^{m,s} \) is called the equivalent depth (Taylor, 1932). Solutions of Laplace’s tidal equation were first explored satisfactorily by Hough (1897, 1898). Hence \( \{\Theta_n^{m,s}\} \) are called Hough functions. Laplace’s tidal equation were solved by Dikii (1965, 1967) and Golstyn and Dikii (1966). Flattery (1967) and Longuett-Higgins (1967) published comprehensive tabulation of the tidal modes. For each \( h_n^{m,s} \) the vertical structure equation (6.9) is solved and tidal fields are computed for each given value of \( h_n^{m,s} \). Kato (1966) and Lindzen (1967) showed that tidal modes of diurnal period could exist with negative values of equivalent depth. Such modes are damped rapidly as they propagate away from generating region and are known as evanescent modes. These modes will have considerable amplitude in the generating region and influence other atmospheric processes in that region.
Equation (6.10) has unique solutions. Given two boundary (upper and lower) conditions, the vertical structure of a given Hough mode is given by equation (6.10) and is called vertical structure equation. Equation (6.9) can be solved for given \( m \) which gives 'equivalent depth' of the atmosphere \( 'h_n' \) and substituting \( h_n \) in equation (6.10), \( L_n \) can be determined to get the height structure function of a particular latitudinal mode. The vertical structure equation has been analysed by Pekeris (1937), Siebert (1961), Buttler and Small (1963) and Lindzen (1967).

6.2.1. Structure of Tidal Perturbation Fields

The tidal perturbation fields are \( P', \rho', T', u, v \) and \( w \) in pressure, density, temperature, southward velocity, eastward velocity and vertical velocity respectively. The horizontal and vertical structure of these tidal fields can be derived as follows (Chapman and Lindzen, 1970).

\[ P' = \sum_n P_n^{m,s}(x) \Theta_n^{m,s}(\theta) \]  
\[ T' = \sum_n T_n^{m,s}(x) \Theta_n^{m,s}(\theta) \]  
\[ w = \sum_n w_n^{m,s}(x) \Theta_n^{m,s}(\theta) \]

and \( u \) and \( v \) expanded as

\[ u = \sum_n u_n^{m,s}(x) U_n^{m,s}(\theta) \]  
\[ v = \sum_n v_n^{m,s}(x) V_n^{m,s}(\theta) \]

where \( P_n^{m,s}, T_n^{m,s}, w_n^{m,s} \) represents the respective vertical structure functions.

\( \Theta_n^{m,s} \) is the Hough function corresponding to a particular eigen value (for fixed \( m \) and \( s \)).

\[ U_n^{m,s}(\theta) = \frac{1}{f^2 - \cos^2 \theta} \left( \frac{d}{d \theta} + \frac{s \cot \theta}{f} \right) \Theta_n^{m,s}(\theta) \]  
\[ V_n^{m,s}(\theta) = \frac{1}{f^2 - \cos^2 \theta} \left( \frac{\cos \theta}{f} \frac{d}{d \theta} + \frac{s}{\sin \theta} \right) \Theta_n^{m,s}(\theta) \]
\( \Theta_n^{m,s} \), \( U_n^{m,s} \) and \( V_n^{m,s} \) represent the horizontal structure functions in the expansion of the tidal fields. \( U_n^{m,s} \) and \( V_n^{m,s} \) are called velocity expansion functions.

\( u_n^{m,s}, v_n^{m,s}, w_n^{m,s}, P_n^{m,s} \) and \( T_n^{m,s} \) are given by

\[
\begin{align*}
\frac{\partial}{\partial x} \left( \frac{d L_n^{m,s}}{dx} - P_n^{m,s} \right) = i g n h_n^{m,s} \left( \frac{d h_n^{m,s}}{dx} + \left( \frac{H}{h_n^{m,s}} - 1 \right) L_n^{m,s} \right) - \frac{i \sigma \Omega_n^{m,s}}{g} \\
\frac{\partial}{\partial z} \left( \frac{d L_n^{m,s}}{dx} - L_n^{m,s} \right) - \frac{1}{g} \Omega_n^{m,s} \\
\frac{\partial}{\partial x} \left( \frac{d L_n^{m,s}}{dx} - L_n^{m,s} \right) - \frac{1}{g} \Omega_n^{m,s} \\
\frac{\partial}{\partial y} \left( \frac{d L_n^{m,s}}{dx} - L_n^{m,s} \right) - \frac{1}{g} \Omega_n^{m,s} \\
\end{align*}
\]

where \( \Omega \) is the gravitational potential of tidal field and \( \bar{p}_0(0) \) is the mean surface pressure of the air.

Looking at equations (6.11) to (6.22) it is seen that in order to determine the tidal fields completely in addition to a knowledge of the source functions \( J_n^{m,s} \) and \( \Omega_n^{m,s}, L_n^{m,s} \) and \( \Theta_n^{m,s} \) must be known for all values of \( n \) and for fixed values of \( m \) and \( s \). For migrating diurnal tides \( m = \omega \) and \( s = 1 \) and for migrating semidiurnal tides \( m = 2 \omega \) and \( s = 2 \).

First Laplace's tidal equation is solved to determine all values of the equivalent depths \( h_n^{m,s} \) and the corresponding Hough functions \( \Theta_n^{m,s} \) for given fixed values of \( m \) and \( s \). Normalised values of \( \Theta_n^{m,s} \) are used in the expansion of tidal fields in equations (6.11) - (6.17).
For each $h_n^{m,s}$ the vertical structure equation is solved knowing the corresponding $J_n^{m,s}$ and $\Omega_n^{m,s}$ and the vertical structure of $H$ and Newtonian Cooling rate coefficient ($\alpha$). Now the tidal fields are readily computed for each eigenvalues $h_n^{m,s}$ from equations (6.18)-(6.22).

6.3. Nomenclature of Tides

For representing migrating tides the form $\Theta_{m,n}$ or $\Theta(m,n)$ is used where $m$ denotes the frequency of tidal oscillations (cycles/day) and $n$ the latitudinal index. Thus for diurnal modes $m = 1$ and semidiurnal modes $m = 2$. Positive values of $n$ correspond to vertically propagating modes and negative values of $n$ correspond to vertically trapped or evanescent modes. The value of $n$ will be positive for gravitational modes and negative for rotational modes. Further the value of $n$ will be fixed in such a way that $(n-m+2)$ gives the number of zeroes in the diurnal gravitational symmetric and antisymmetric modes; $(|n|-m-1)$ in the diurnal rotational symmetric modes; $(|n|-m+1)$ in the diurnal rotational antisymmetric modes and $(n-m)$ in the semidiurnal gravitational symmetric and antisymmetric modes. Thus diurnal gravitational symmetric Hough functions are designated by $\Theta(1, 1), \Theta(1, 3), \Theta(1, 5)$... diurnal gravitational antisymmetric Hough functions by $\Theta(1, 2), \Theta(1, 4), \Theta(1, 6)$... diurnal rotational symmetric Hough functions by $\Theta(1, -2), \Theta(1, -4), \Theta(1, -6)$... diurnal rotational antisymmetric Hough functions by $\Theta(1, -1), \Theta(1, -3), \Theta(1, -5)$... semidiurnal gravitational symmetric Hough functions by $\Theta(2, 2), \Theta(2, 4), \Theta(2, 6)$... and semidiurnal gravitational antisymmetric Hough functions by $\Theta(2, 3), \Theta(2, 5), \Theta(2, 7)$... .

For representing nonmigrating tidal modes, one more index ‘$s$’ is used in addition to $m$ and $n$ i.e. the triplet notation $(m,s,n)$ is used to specify the eigen value/ eigen vector. Each triplet defines an eigen value/eigen vector problem that uniquely determines an equivalent depth and Hough function i.e. each triplet defines a ‘mode’. ‘$s$’ is called the zonal wave number. The zonal wave number, $s$, defines the longitudinal structure of the tidal oscillation. The absolute value of $s$ defines how many complete sine waves of the oscillation span the globe latitudinally. The zonal wave number must be an integer to insure continuity of the longitudinal structure.
Positive (negative) zonal wave numbers describe longitudinal structure with phase progressing westward (eastward) with time. The tides can not follow the motion of the sun around the globe in the zonal direction, if ‘m’ is not equal to ‘s’. Thus in general designation tidal modes \( m = s \) represents migrating component and ‘m’ not equal ‘s’ represents nonmigrating components.

6.4. Simulation of Tidal Fields over Gadanki and Comparison with Observations

Migrating diurnal tides are primarily generated by solar radiation absorption by water vapour (Chapman and Lindzen, 1970) and nonmigrating tides mainly by other diurnal sources such as planetary boundary layer heat flux and latent heat release in clouds. A recent study by Williams and Avery (1996a) of the diurnal deep convective activity (DCA) based on infrared radiance measured by four geostationary and two polar orbiting satellites shows that westward propagating nonmigrating modes with zonal wave number 5 and eastward propagating nonmigrating modes with zonal wavenumber -3 are dominantly present (in addition to the zonal wave number 0) in the two equinox seasons and winter season. For summer season the zonal wave number distribution of the diurnal DCA is distinctly different from the other three seasons. For summer season, only zonal wavenumber 5 is present. Tsuda and Karo (1989) has shown a similar zonal wave number distribution with the dominance of +5 and -3 zonal wave numbers for the diurnal planetary boundary layer heat flux. Thus it appears that diurnal heating due to both latent heat release in deep convective clouds and planetary boundary layer heat flux are strongly linked to the land-sea distribution as noted by Williams and Avery (1996a). For this reason, in the present simulation only zonal wave numbers +5 and -3 were used for both latent heat release and planetary boundary layer heating. Further, for planetary boundary layer heat flux one westward propagating nonmigrating mode \((1, 5, 16)\) with a short vertical wavelength of \( ^{-3} \) km was included, considering the narrow vertical extent of the diurnal heating region. For similar reasons this mode was included for diurnal cloud heating which is expected to have a very narrow heating region in the vertical direction. For the two equinox seasons, only one simulation was done using diurnal heating rates for different heat
sources. For winter and summer separate simulations were done changing the nonmigrating modes, which are dominantly present for each season.

The latent heat source drives migrating diurnal modes also, which are weak when compared to the nonmigrating modes (Williams and Avery, 1996 a). A similar argument is valid for PBL also (e.g. Tsuda and Kato, 1989). In the present work it is assumed that for diurnal solar short-wave heating of clouds also a dominance of nonmigrating modes exists. Similarly, water vapour drives nonmigrating modes also. It is known that diurnal nonmigrating modes generated by water vapour heating is very weak when compared with the (1,1,1) migrating mode which is stronger by an order of magnitude (e.g. Groves, 1982). As far as ozone diurnal heating is concerned the large vertical extent of the heating region (20-80 km) causes destructive interference of the short-wavelength diurnal modes and is practically ineffective in contributing to the diurnal fields. In the present simulation of diurnal tides, therefore, only migrating modes due to water vapour heating, nonmigrating modes due to PBL heating, latent heat and solar short-wave heating of clouds are included.

As discussed above, for the present simplified simulation, only migrating modes (1, 1, 1), (1, 1, -2), (1, 1, -4), (1, 1, -6), (1, 1, -1), (1, 1, -3), (1, 1, 3) and (1, 1, 2) were used for water vapour heating and the heating rates taken from Groves (1982). The vertical profile of the heating rate due to water vapour is shown in figure 6.1. For planetary boundary layer heating, westward and eastward travelling nonmigrating modes (1, 5, 5), (1, 5, 6), (1, 5, 7), (1, 5, 8), (1, 5, 10), (1, 5, 16), (1, -3, 3), (1, -3, 4) and (1, -3, 5) were used. The vertical structure of the heating rates due to planetary boundary layer heat flux is shown in figure 6.2. The vertical structure of the heating rates of all these modes corresponding to the diurnal planetary boundary layer heating had an e-folding depth of ~1.5 km with the maximum occurring at the surface. The heating rate values at the surface varied from 125 mW kg\(^{-1}\) (for 1, 5, 10 mode) to 10 mW kg\(^{-1}\) (for 1, -3, 5 mode) for the different modes. For heating rates due to latent heat release, nonmigrating modes (1, 5, 5), (1, 5, 6), (1, 5, 7), (1, 5, 8), (1, 5, 10), (1, -3, 3), (1, -3, 4) and (1, -3, 5) were used. The vertical profile of the heating rate due to latent heat release is shown in figure 6.3. The vertical extent of the heating region extended from ~1 to ~14 km.
Figure 6.1. Heating rate due to water vapour
Figure 6.2. Heating rate due to planetary boundary layer heat flux
Figure 6.3. Heating rate due to latent heat release
with a broad maximum at \( -9 \) km. The maximum heating rate values varied from \( -7 \text{ mW kg}^{-1} \) (for 1.5.10 mode) to 0.7 \text{ mW kg}^{-1} \) (for 1. -3, 5 mode) for different modes. For heating rates due to cloud heating, nonmigrating modes (1, 5, 5), (1, 5, 6), (1, 5, 7), (1, 5, 8), (1, 5, 10), (1, 5, 16), (1, -3, 3), (1, -3, 4) and (1, -3, 5) were used for winter season since all the modes are stronger during winter. For the cloud heating only two nonmigrating modes were considered for equinox season namely, (1, -3, 5) and (1, 5, 16) with a narrow peak centred at \( -8 \) km. The e-folding depth of the heating region was \( -2 \) km with maximum values varying from \( -80 \text{ mW kg}^{-1} \) (for 1,5,16 mode) to 10 \text{ mW kg}^{-1} \) (for 1, -3, 5 mode). For summer, only zonal wavenumber 5 were used. The vertical profile of heating rate due to cloud heating is shown in figure 6.4. The nonmigrating modes (1, 5, 5), (1, 5, 6), (1, 5, 7), (1, 5, 8), (1, 5, 10), (1, 5, 16) are used for planetary boundary layer heating, heating due latent heat release and cloud heating during summer season. The Hough modes and their respective equivalent depths used for different sources are summarised in Table 6.1.

These above Hough modes corresponding to different diurnal heat sources were used for computing the tidal response of the lower atmosphere up to 20 km altitude using classical tidal theory. Horizontal velocity fields \( u \) (southward) and \( v \) (eastward) are expanded using velocity expansion functions \( U_{n}^{m,n}(\theta) \) and \( V_{n}^{m,n}(\theta) \) given by equations (6.16) and (6.17). Then vertical structure equation (6.10) is solved using Newtonian cooling coefficients \( \alpha(z) \) and the heating rates \( J_{n}^{m,n} \) due to water vapour, planetary boundary layer heat flux, latent release from deep convective clouds and cloud forcing. The total fields \( u \) and \( v \) are computed using equations (6.14) and (6.15).

The classical theory of tides, which is used for the simulation of diurnal tides, has many limitations. This theory assumes that the atmosphere is motionless. The zonal mean winds present in the atmosphere can affect the propagation of the diurnal tidal oscillations through it by changing the phase velocity of the waves. However, the mean zonal winds are generally very weak in the tropical lower atmosphere except during June to September period when tropical easterly jet below
Figure 6.4. Heating rate due to cloud forcing
### Table 6.1

The Hough modes and the corresponding equivalent depths.

<table>
<thead>
<tr>
<th>Source</th>
<th>$\Theta$ (m, s, n)</th>
<th>$h$ (m, s, n)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Water Vapour Heating</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(1, 1, 1)</td>
<td></td>
<td>0.691</td>
</tr>
<tr>
<td>(1, 1, -2)</td>
<td></td>
<td>-12.154</td>
</tr>
<tr>
<td>(1, 1, -4)</td>
<td></td>
<td>-1.7398</td>
</tr>
<tr>
<td>(1, 1, -6)</td>
<td></td>
<td>-0.6374</td>
</tr>
<tr>
<td>(1, 1, -1)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>(1, 1, -3)</td>
<td></td>
<td>-1.7941</td>
</tr>
<tr>
<td>(1, 1, 3)</td>
<td></td>
<td>0.1209</td>
</tr>
<tr>
<td>(1, 1, 2)</td>
<td></td>
<td>0.2397</td>
</tr>
<tr>
<td><strong>Heating due to Planetary Boundary Layer Heat Flux</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(1, 5, 5)</td>
<td></td>
<td>0.3581</td>
</tr>
<tr>
<td>(1, 5, 6)</td>
<td></td>
<td>0.1762</td>
</tr>
<tr>
<td>(1, 5, 7)</td>
<td></td>
<td>0.1015</td>
</tr>
<tr>
<td>(1, 5, 8)</td>
<td></td>
<td>0.0650</td>
</tr>
<tr>
<td>(1, 5, 10)</td>
<td></td>
<td>0.0328</td>
</tr>
<tr>
<td>(1, 5, 16)</td>
<td></td>
<td>0.0092</td>
</tr>
<tr>
<td>(1, -3, 3)</td>
<td></td>
<td>2.2493</td>
</tr>
<tr>
<td>(1, -3, 4)</td>
<td></td>
<td>0.7820</td>
</tr>
<tr>
<td>(1, -3, 5)</td>
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<td>0.3325</td>
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<tr>
<td><strong>Heating due to Latent Heat Release</strong></td>
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<td></td>
<td>0.3581</td>
</tr>
<tr>
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</tr>
<tr>
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<tr>
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<td>0.0650</td>
</tr>
<tr>
<td>(1, 5, 10)</td>
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<tr>
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<td>2.2493</td>
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<td>0.7820</td>
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<tr>
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<td><strong>Heating due to Cloud Forcing</strong></td>
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<td>(1, 5, 7)</td>
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<td>(1, -3, 5)</td>
<td></td>
<td>0.3325</td>
</tr>
</tbody>
</table>
the tropopause can attain magnitudes as large as 25-30 m s\(^{-1}\). For the migrating and nonmigrating modes which we have included in the present simulation of diurnal tides, the propagation of the tidal oscillations are not affected by the mean winds due to the large phase speed of these oscillations ranging from \(\sim 500\) to \(\sim 100\) m s\(^{-1}\). Therefore, simulation of diurnal tidal fields including the above mentioned modes using classical theory will yield reasonable results in the tropical lower atmosphere. However, if larger zonal wave number modes are included then the effects of mean winds on the tidal fields have to be considered. Another assumption is the independence of different Hough modes whereas in a real atmosphere mode-coupling is observed in which case the modes not present in the excitation sources are generated. But the classical theory cannot take into account this mode-coupling. However, this process may be neglected in the lower atmospheric region.

The vertical structure of tidal response of combinations of various nonmigrating heat sources along with water vapour that are used in the simulations is shown in figures 6.5 to 6.10. In these figures the simulated responses are compared with observed vertical structure of diurnal tides over Gadanki during autumnal equinox season. The circles represent the simulated values and triangles represent the observed values.

Figure 6.5 shows the tidal response when the heating rates due to water vapour (migrating) and planetary boundary layer heat flux (nonmigrating) only are considered. The amplitude values are slightly larger than the observed values except at few altitudes for both zonal and meridional wind. The phase structure of meridional wind of both simulated and observed ones are similar in the altitude region of 12-20 km. The phase structure of meridional wind of simulated and observed values are very much differing in the 8-12 km altitude region. For zonal wind simulated and observed phase structure are very much dissimilar except at lower altitudes (3-8 km). Between 14-18 km the phase structure of zonal wind of simulated and observed values are similar with a phase difference of \(\sim 4\) hours. Similarly between 10-14 km phase structure is similar with a phase difference of \(\sim 8\)
Figure 6.5. Vertical profiles of amplitude (left panel) and phase (right panel) of diurnal tides in meridional (top) and zonal (bottom) winds observed over Gadanki (open triangles) during autumnal equinox season. Amplitudes and phases simulated by classical tidal theory using heating rates due to water vapour and planetary boundary layer heat flux for the location of Gadanki are also shown (filled circles). The r.m.s. error are shown as horizontal bars at every ~1 km.
hours for zonal winds. These dissimilarities shows that for simulation some other heat sources also should be included.

Figure 6.6 shows the tidal response when the heating rates due to water vapour (migrating) and latent heat release in deep convective clouds (nonmigrating) only are considered. Though diurnal amplitudes of the meridional wind show good agreement with the observed ones in limited height regions (3-8 km and 14-19 km), in general, there exist large discrepancies between the simulated and observed diurnal phases. For the zonal winds of diurnal tides, the observed and simulated amplitudes are similar in the 5-10 km and 13-19 km altitude region. The phase structures of observed and simulated diurnal zonal winds are dissimilar. It is inferred that heating due to latent heat release along with water vapour alone could not account for the observed tidal response during autumnal equinox season.

Figure 6.7 shows the tidal response when the heating rates due to water vapour (migrating) and cloud forcing (nonmigrating) are considered. In the lower atmosphere up to 10 km the altitude structure of amplitude of both observed and simulated values are very much matching for both meridional and zonal winds. For meridional wind simulated amplitude values are greater than the observed ones above 10 km. The simulated values of amplitudes are greater than observed values above 12 km for zonal winds. Above 14 km both simulated and observed phase structures show downward phase propagation for meridional diurnal winds. There is a phase difference of ~4 hours between the two. For meridional winds the phase structure of simulated and observed values are differing much, below 14 km. The simulated phase structure of zonal winds is similar to the observed ones above 15 km with a phase difference of ~8 hours. Below 15 km the phase structure of zonal winds could not be reproduced in the simulated structure.

Figure 6.8 shows the observed amplitude and phase structure of both zonal and meridional winds during autumn equinox season and simulated ones obtained using the heating rates due to water vapour, planetary boundary layer heat flux and latent heat release in deep convective clouds. The amplitude structure of meridional
Figure 6.6. Vertical profiles of amplitude (left panel) and phase (right panel) of diurnal tides in meridional (top) and zonal (bottom) winds observed over Gadanki (open triangles) during autumnal equinox. Amplitudes and phases simulated by classical tidal theory using heating rates due to water vapour and latent heat release for the location of Gadanki are also shown (filled circles). The r.m.s. error are shown as horizontal bars at every -1 km.
Figure 6.7. Vertical profiles of amplitude (left panel) and phase (right panel) of diurnal tides in meridional (top) and zonal (bottom) winds observed over Gadanki during autumnal equinox. Amplitudes and phases simulated by classical tidal theory using heating rates due to water vapour and cloud forcing for the location of Gadanki are also shown (filled circles). The r.m.s. error are shown as horizontal bars at every ~1 km.
Figure 6.8. Vertical profiles of amplitude (left panel) and phase (right panel) of diurnal tides in meridional (top) and zonal (bottom) winds observed over Gadanki (open triangles) during autumnal equinox. Amplitudes and phases simulated by classical tidal theory using heating rates due to water vapour, planetary boundary layer heat flux and latent heat release for the location of Gadanki are also shown (filled circles). The r.m.s. error are shown as horizontal bars at every ~1 km.
wind of observed and simulated values are similar in the 3-13 km height region with simulated ones being slightly larger. For zonal winds the amplitude structure of observed and simulated values are similar in the 3-10 km and 12-17 km height region. The amplitude peak observed for zonal winds at 10 km could not be reproduced in the simulated structure. The phase structures of observed and simulated ones shows upward propagation in the 3-6 km and 9-14 km altitude regions for meridional winds. The phase structures of zonal winds of observed and simulated ones are very much dissimilar.

Figure 6.9 shows the observed amplitude and phase structure of both zonal and meridional winds during autumn equinox season and simulated ones obtained using the heating rates due to water vapour, planetary boundary layer heat flux and cloud forcing. For meridional winds the simulated amplitudes of diurnal oscillations are larger than the observed ones. The amplitude values obtained in the simulation for zonal winds are similar to those observed in the 3-10 km and 15-18 km height regions. For zonal winds the amplitude peak observed at 10 km is not obtained in the simulated structure. The phase structures of observed and simulated ones show upward propagation in the 3-6 km and 7-12 km altitude regions for meridional winds. The observed phase structure of zonal winds could not be reproduced in the simulated structure.

The observed and simulated (using heating rates due to water vapour, latent heat release in deep convective clouds and cloud forcing) altitude structures of amplitude and phase of zonal and meridional components of diurnal oscillations are shown in figure 6.10. The simulated amplitude structure is similar to observed one in the lower atmosphere up to 10 km for meridional winds. For meridional winds the simulated amplitudes are larger than the observed ones. The simulated amplitudes are matching with observed ones for zonal winds up to 10 km. Between 10 and 12 km the simulated values are less than the observed values for zonal winds and above 12 km the simulated values are slightly larger than the observed ones. The phase structure of both observed and simulated meridional winds are similar above 12 km altitude with a phase difference of ~4 hours between the two. Below 12
Figure 6.9. Vertical profiles of amplitude (left panel) and phase (right panel) of diurnal tides in meridional (top) and zonal (bottom) winds observed over Gadanki (open triangles) during autumnal equinox. Amplitudes and phases simulated by classical tidal theory using heating rates due to water vapour, planetary boundary layer heat flux and cloud forcing for the location of Gadanki are also shown (filled circles). The r.m.s. error are shown as horizontal bars at every ~1 km.
Figure 6.10. Vertical profiles of amplitude (left panel) and phase (right panel) of diurnal tides in meridional (top) and zonal (bottom) winds observed over Gadanki (open triangles) during autumnal equinox. Amplitudes and phases simulated by classical tidal theory using heating rates due to water vapour, latent heat release and cloud forcing for the location of Gadanki are also shown (filled circles). The r.m.s. error are shown as horizontal bars at every ~1 km.
km the phase structure of observed and simulated meridional winds are entirely different. The simulated phase structure of zonal winds is similar to the observed ones above 15 km with a phase difference of ~8 hours. Below 15 km the phase structure of zonal winds could not be reproduced in the simulated structure.

From figures 6.5 to 6.10 it is inferred that the simulations done using different heat sources (nonmigrating) separately with water vapour (migrating) or combining two of them with water vapour could not reproduce all the features of diurnal tides observed over Gadanki. The results of these studies shows that many observed features of meridional and zonal wind amplitudes are reproduced when cloud forcing was included in the simulation along with water vapour (figure 6.7) and water vapour and latent heat release (figure 6.10) in the 3-10 km height region. In the limited height regions (15-18 km and 14-18 km for zonal winds) both the simulated and observed phase structures also show quite good agreements. The combination of water vapour and planetary boundary layer (figure 6.5) gives simulated amplitudes and phases is not as good as in the case of cloud forcing was included along with water vapour and latent heat release. Though diurnal amplitudes of the meridional wind simulated using water vapour and latent heat release show good agreement with the observed ones in limited height regions (3-7 km and 14-19 km) (figure 6.6), in general, there exist large discrepancies between the simulated and observed diurnal phases. The phase structure obtained when planetary boundary layer heat flux and latent heat released are used along with water vapour for simulation show good agreement with observed ones in the 3-5 and 8-13 km altitude region for meridional winds (figure 6.8). Similar phase structure is obtained when planetary boundary layer heat flux and cloud forcing was used along with water vapour for simulation (figure 6.9). From above observations it is inferred that some of combinations of the above heat sources are important at certain height regions in producing diurnal tides at those height regions. So it is important to simulate the diurnal tides including all possible heat sources.
Now the simulation of diurnal tides is done using the heating rates due to all the four sources. The possible seasonal variations of the sources are also taken into account. Nonmigrating modes with the zonal wave number 5 and -3 are dominantly present in the two equinox seasons and winter season (Williams and Avery 1996a) over the equatorial region. For summer nonmigrating modes with zonal wave number 5 only is present. For the present simulation only zonal wave number 5 and 3 are used during equinoxes and winter. During equinoxes for the cloud forcing only two nonmigrating modes were considered namely (1, -3, 5) and (1, 5, 16) with a narrow peak centred at -8 km. During winter all the nonmigrating modes are stronger. So for cloud forcing nine strongest nonmigrating modes are used. For summer season the zonal wavenumber 5 only is considered for all the nonmigrating sources. The simulated altitude structure of amplitude and phase of zonal and meridional diurnal oscillations are compared with the observed structures for all the four seasons which are shown in figures 6.11 to 6.14.

Figure 6.11 shows the vertical structure of the amplitudes and phases of both the observed and simulated diurnal oscillation in meridional and zonal winds respectively during autumn equinox season. The r.m.s. errors in the amplitudes and phases at every -1 km intervals are shown by horizontal bars. The observed amplitudes are, in general, smaller below about -9 km height with values less than 1 m s⁻¹ for both meridional and zonal components. Both amplitudes and phases show vertical structures indicating interference of different tidal modes. It is seen that for the meridional component both the observed and simulated amplitudes and phases show vertical structures which are very similar. Though the absolute values of the observed and simulated amplitudes differ, some of the maxima and minima in amplitude structure are reproduced. For example, amplitude minima near 6 and 12 km and maximum near 16 km are present in both the profiles. The phase structure in the entire height region is almost reproduced in the simulated profiles. Above 8 km height, observed phase values closely follow the simulated ones. For the zonal component the amplitude maxima near 5 and 8 km and minimum near 16 km are reproduced in the simulated profile also. The observed and simulated phase profiles of zonal component differ very much unlike the phase profile of the meridional
Figure 6.11. Vertical profiles of amplitude (left panel) and phase (right panel) of diurnal tides in meridional (top) and zonal (bottom) winds observed over Gadanki (open triangles) during autumn equinox season. Amplitudes and phases simulated by classical tidal theory for the location of Gadanki are also shown (filled circles). The r.m.s. error are shown as horizontal bars at every ~1 km.
component, though in a limited altitude region the two profiles agree with each other.

The diurnal amplitudes and phases of the meridional and zonal components of the winds in the winter season are given in figure 6.12. For the meridional component in winter season, observed amplitudes show 3 maxima near 7, 11 and 16 km altitudes and the simulated profile is able to reproduce only 2 maxima near 7 and 16 km. In winter also the values of the observed and simulated amplitudes differ very much, the simulated ones being greater. The phase structure in the entire height region is almost reproduced in the simulated profiles. There is a small phase difference between observed and simulated phase structure in the 8-12 km altitude region. The zonal component in winter season shows large differences between the observed and simulated amplitude and phase structure. In the case of zonal winds also the simulated amplitudes are, in general, larger than the observed ones. The observed phase structure in the 7-12 km height region shows a clear downward propagation with a vertical wavelength of \( \sim 6 \) km whereas the simulated phase profile shows a much larger vertical wavelength.

The observed vertical amplitude structures of diurnal tides in the autumn (figure 6.11) and vernal equinox (figure 6.13) seasons are almost similar with the amplitudes in the vernal equinox season being larger. The amplitude maxima near 5 and 10 km are present during both the seasons, the amplitude being very much larger during the vernal equinox season. The observed phase structures during the two seasons (figure 6.11 and figure 6.13) are almost similar except for a phase difference between the two. In both the equinox seasons the general shape of the observed vertical structures is very similar to the simulated ones. Two phase reversal regions, one near 13 km and second one near 8 km can be clearly seen in both the observation and in the simulation during the equinox seasons. But, in the case of vernal equinox season there is a large phase difference between the observed and the simulated one in the phase reversal region. Another interesting feature during vernal equinox season is the downward phase propagation with a vertical wavelength of \( \sim 3 \) km in the 17-20 km height region and \( \sim 6 \) km in 14-17 km height.
Figure 6.12. Vertical profiles of amplitude (left panel) and phase (right panel) of diurnal tides in meridional (top) and zonal (bottom) winds observed over Gadanki (open triangles) during winter season. Amplitudes and phases simulated by classical tidal theory for the location of Gadanki are also shown (filled circles). The r.m.s. error is shown as horizontal bars at every ~1 km.
Figure 6.13. Vertical profiles of amplitude (left panel) and phase (right panel) of diurnal tides in meridional (top) and zonal (bottom) winds observed over Gadanki (open triangles) during vernal equinox season. Amplitudes and phases simulated by classical tidal theory for the location of Gadanki are also shown (filled circles). The r.m.s. error are shown as horizontal bars at every ~1 km.
region (figure 6.13). But for the zonal components (figures 6.11 and 6.13) both the amplitudes and phases show large differences between the two seasons. The zonal wind amplitudes during the vernal equinox season are, in general, larger than those during autumn equinox season. During vernal equinox season the 10-17 km shows nearly a constant phase region and a clear downward phase propagation with a vertical wavelength of ~4 km is observed below 8 km height region.

The observed diurnal amplitudes and phases of both meridional and zonal components in summer are significantly different from those during other three seasons (figure 6.14). The observed diurnal meridional component shows prominent amplitude maximum near 7 km, 10 km, and 16 km heights during the summer (figure 6.14) season. The amplitude maxima at 7 km and 16 km altitudes are reproduced in the simulated profile also. For the meridional diurnal wind phase there is good agreement between the observed and simulated profiles in the entire height region. For zonal component the observed and simulated amplitude structure are almost similar up to ~16 km. The phase structure of zonal component of observed and simulated ones differs much, except at few altitude regions.

One common characteristic observed during all four seasons is the large discrepancy between the simulated and observed phase structures of the diurnal zonal wind component. The exact reason for this discrepancy is not clearly identified. However, dominance of different heat sources during different seasons could be a plausible reason for this. For example, observed zonal wind phase structure during vernal equinox season could be reproduced reasonably well (figure 6.15) when only water vapour and latent heat were included in the simulation. The simulated phase profile could reproduce the almost constant phase region above ~10 km and downward phase propagation region below ~7 km heights.

There is apparently fairly good agreement between the observed and simulated diurnal tidal oscillation of the meridional wind during all the four seasons, especially in the phase structure. It appears that the diurnal fields in the troposphere and lower stratosphere are produced as a result of superposition of many modes.
Figure 6.14. Vertical profiles of amplitude (left panel) and phase (right panel) of diurnal tides in meridional (top) and zonal (bottom) winds observed over Gadanki (open triangles) during summer season. Amplitudes and phases simulated by classical tidal theory for the location of Gadanki are also shown (filled circles). The r.m.s. error are shown as horizontal bars at every ~1 km.
Figure 6.15. Vertical profiles of amplitude (left panel) and phase (right panel) of diurnal tides in zonal winds observed over Gadanki (open triangles) during vernal equinox season. Amplitudes and phases simulated by classical tidal theory using heating rates due to water vapour and latent heat release for the location of Gadanki are also shown (filled circles). The r.m.s. error are shown as horizontal bars at every ~1 km.
both migrating and nonmigrating, generated by various sources such as water vapour heating, planetary boundary layer heating, latent heat release and solar short-wave radiation absorption by clouds. The classical theory of tides is reasonably successful in simulating the diurnal tidal structure in the lower atmosphere, especially during autumnal equinox season and winter season (to a lesser extent). The height region of phase reversal, in the mid-troposphere near 9 km could be simulated well, only when the cloud heating was included. This shows the importance of solar short-wave heating of clouds in generating diurnal tidal oscillations in the lower atmosphere, similar to the findings of Braswell and Lindzen (1998) in the case of diurnal surface pressure oscillation.

6.5. Conclusion

The amplitude maxima and minima obtained at different heights in the observed vertical profile of amplitudes, in general, show close similarities with amplitude profiles simulated in the present study, which suggests that these are produced as a result of interference of many Hough modes generated by the above mentioned heat sources. Similarly there appears to be a gross agreement between the vertical profile of the observed and simulated phases of the diurnal component in meridional wind during all the four seasons which is suggestive of interference of different Hough modes in the troposphere and lower stratosphere. Another observation is that though the vertical profile of diurnal phase shows upward propagation in certain height regions, it need not necessarily correspond to downward propagation of tidal energy; they are primarily the manifestations of interference between different Hough modes. But there is also a complex situation arising from the propagation of waves from the planetary boundary layer into the mid-troposphere where other tidal sources are primarily situated. The phase reversals in the mid-troposphere are simulated well, only when the cloud heating was included. This shows the importance of solar short wave heating of clouds in the lower atmosphere.