Variability of mesospheric diurnal tides and tropospheric diurnal heating during 1997–1998

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This study focuses on interannual variations of diurnal tropospheric heating and the response in the mesosphere observed by radars and predicted by a model. The work is prompted by reports of interannual variability in amplitudes of tidal variables at low latitudes. Diurnal tides observed at Hawaii and Christmas Island exhibit a pronounced “spike” in amplitude from late 1997 to early 1998. It has been speculated that this variability may be linked to the El Niño–Southern Oscillation phenomenon. We examine diurnal solar heating due to water vapor absorption, and diurnal latent heat release due to deep convection between 1988 and 2005. Both of these heating drives exhibit anomalously higher amplitudes in the tropical central and eastern Pacific during 1997–1998. The altered heating patterns result in a stronger forcing of the migrating diurnal tide by water vapor heating, and excitation of several weaker nonmigrating modes by latent heating. A primitive equation model is used to evaluate how these drives contribute to diurnal winds in the mesosphere. Anomalous water vapor heating results in about 15% increases in model meridional wind amplitudes over climatological values at subtropical latitudes between 300°E and the Greenwich meridian. While the timing of the model amplitude enhancements is consistent with observations at Hawaii, the observed increases are significantly stronger. Our study indicates that water vapor heating is the larger contributor to tidal enhancement observed during 1997–1998.


1. Introduction

Atmospheric tides are among the most prominent motions in the mesosphere and lower thermosphere (MLT), often dominating the meridional wind field at low latitudes [Hays et al., 1994]. Diurnal amplitudes in the MLT exhibit variations on timescales ranging from days to years [Fritts and Isler, 1994; Burrege et al., 1995; Eckermann et al., 1997; Nakamura et al., 1997; Fritts et al., 1997; Pancheva et al., 2002]. The possible causes have not been fully sorted out, but are generally thought to fall into two categories: (1) variability of tropospheric or stratospheric tidal forcing and (2) amplitude modulation due to interaction with the mean flow [Forbes and Vincent, 1989; Hagan et al., 1999; McLandress, 2002b], planetary waves [Manson et al., 1982; Teitelbaum and Vial, 1991; Palo et al., 1998; Palo et al., 1999; Pancheva et al., 2002; Lieberman et al., 2004] and gravity waves [Walterscheid, 1981; Fritts and Vincent, 1987; McLandress and Ward, 1994; Nakamura et al., 1997; Meyer, 1999; McLandress, 2002a].

This study addresses variability in diurnal tropospheric heating mechanisms, and its effects upon MLT tides. The work is prompted by reports of interannual (IA) variability in tropospheric and mesospheric tidal amplitudes. We examine two primary sources of diurnal tides: heating due to water vapor insolation absorption and latent heat release in deep convection. We focus on events during 1997–1998, that are associated with the El Niño–Southern Oscillation (ENSO). ENSO is a global-scale perturbation in atmospheric pressure, cloudiness, temperature and rainfall that occurs roughly every 2–10 a, when warm sea surface temperatures (SST) are displaced from the western into the central and eastern tropical Pacific [Rasmusson and Wallace, 1983; Philander, 1990]. The resulting variations in large-scale tropical weather and climate systems may affect the climateology of the middle atmosphere through modulation of the radiative/convective excitation of vertically propagating waves: tides, Kelvin waves, and gravity waves [Vial et al., 1994; Gage et al., 1996; Eckermann et al., 1997; Mote et al., 1998; Mote et al., 2000; Gavrilov et al., 1999; Thompson and Wallace, 2000; Thompson et al., 2000].

Evidence for IA variability in surface diurnal tides was presented in a study by Vial et al. [1994] that examined multidecadal time series of the diurnal surface pressure tide at Macao (22°N, 113°E) and Batavia (now Jakarta, 6°S,
They reported IA modulations in diurnal amplitude of about 10–20% of the mean annual amplitude (≈0.5 hPa). The variations were correlated with a Southern Oscillation index (SOI) based upon the Tahiti-Darwin surface pressure difference. Vial et al. [1994] speculated that the pressure variations at Batavia and Macao were manifestations of diurnal forcing by water vapor absorption and latent heat release that have been altered on planetary (or at least Pacific Ocean size) scales by the ENSO phenomenon.

Multiyear studies of the MLT diurnal tide have been presented by Burrage et al. [1995], Vincent et al. [1998], Tsuda et al. [1999], and Gurubaran and Rajaram [1999]. All reported IA variability of the tide that appeared to be related to the stratospheric quasi-biennial oscillation. More recently, Gurubaran et al. [2005] documented variations in MLT diurnal winds during 1993–1999 at Tirunelveli (8.7°N, 78°E) and Jakarta that correlated with deep convective proxy data over the central Pacific. Gurubaran et al. [2005] speculated that ENSO-related tidal variability arose through convective excitation of nonmigrating tides, that competed for dominance with the migrating diurnal tide.

In the present study we quantify IA variations in MLT diurnal tides, and in their radiative and convective forcing in the troposphere. We then use a primitive equation model to evaluate how the driving is communicated to diurnal winds in the mesosphere. These experiments enable us to compare the relative contributions of latent heat release and water vapor insolation absorption to IA MLT tidal variations. Our results confirm previous suggestions that the ENSO phenomenon modifies tidal forcing. The key agent appears to be enhancement of migrating component of the diurnal water vapor insolation absorption. This is a consequence of the increase in specific humidity over the central and eastern Pacific during 1997–1998.

2. Mesospheric Tidal Observations

We begin by showing tidal winds from Kauai, since the variability exhibited at this and other sites was one of the primary motivations for our study. The MF radar on the island of Kauai, Hawaii (22°N, 154°W) has a nearly continuous time series of winds extending over 17 a, probably the longest existing data set that can be used to demonstrate mesospheric tidal variability at tropical latitudes. 17-a diurnal composites were prepared for each month of data, and then sinusoidal functions were fit to provide estimates of zonal and meridional wind amplitudes.

Figure 1 shows the variability of the diurnal tide at 90 km. During periods when the actual tidal amplitude exceeds the composite value, the difference is filled with red. When the tidal amplitude falls below the composite value, the difference between the composite value and the actual value is in blue.

![Figure 1. Diurnal tidal amplitudes as a function of year, at Kauai (22°N, 154°W). When amplitudes exceed the average seasonal value, the area between the seasonal value and the actual value for that time is in red. When amplitudes fall below the seasonal value, the difference between the seasonal value and the actual value is in blue.](image-url)
Island (2°N, 157°W) (see Figure 2). An MF radar at Rarotonga (22°S, 160°W) was not operating during the large 1997–1998 tidal enhancement, but this site is shown to demonstrate that the tidal variability at this southern hemisphere site closely tracks that in the northern hemisphere. Gurubaran et al. [2005] also documented enhancements in mesospheric diurnal winds at Tirunelveli and Jakarta, that occurred in early to mid-1997. Therefore we are convinced that the 1997–1998 variations exhibited in Figures 1 and 2 are real. There appears to be a high correlation in the tidal amplitudes between low-latitude radar sites over India, the maritime continent, and the central Pacific Ocean.

3. Tropospheric Tidal Sources

[10] The key tropospheric drivers for vertically propagating diurnal tides are heating due to absorption of IR solar radiation by water vapor (hereafter referred to as “water vapor heating”), and latent heat release due to deep convection [Chapman and Lindzen, 1970; Hagan, 1996; Hagan and Forbes, 2002]. We have investigated these sources using data from the NASA Water Vapor Project (NVAP), the International Satellite Cloud Climatology Project (ISCCP), and the Tropical Rainfall Measuring Mission (TRMM).

3.1. IR Absorption

[11] Diurnal heating due to IR absorption by tropospheric water vapor was computed using 14 a of monthly mean NVAP global precipitable water (1988–2001). Our methodology is described by Lieberman et al. [2003], where NVAP-based climatological IR heating rates are also presented (using data for 1988–1997). Briefly, NVAP precipitable water values are inverted onto vertical structures of specific humidity provided by concurrent National Center for Earthquake Prediction reanalyses. Specific humidity profiles are then input to a simple two-stream model with vertically uniform cloud layer structure [Groves, 1982].

[12] Monthly composites of the diurnal heating have been formed on the basis of data between 1988 and 2001. These composites are hereafter denoted as “climatology”. Examination of the month-to-month IR diurnal heating rates show that the only significant deviations from climatological conditions occurred between July 1997 and March 1998. The remainder of this discussion focuses upon the heating during this period, compared to the previous and subsequent years.

[13] Diurnal IR heating amplitudes for January 1997, 1998, and 1999 are shown in Figure 3 (left). Figure 3 (right) show the corresponding deviations from climatology. The hour of maximum of the heating (not shown) is uniformly 1200LT. Amplitudes in 1997 (Figure 3, top left) resemble January climatological conditions. The maximum heating anomalies in the eastern Pacific during 1998 are approximately 3 mW kg⁻¹. Comparison with the 14-a climatological heating values (not
indicate that the local anomalies in January 1998 represent a 36% increase over the "typical" heating.

[15] The timing and location of the water vapor IR heating anomaly are broadly consistent with the ENSO phenomenon. During an ENSO event, regions of large-scale tropical moisture and convection shift from the western to the central and eastern Pacific. Figure 4 shows the correlation between monthly tropospherically averaged diurnal IR heating, and the SOI during 1988–2001. Sustained negative values of the SOI are associated with El Niño episodes. The most prominent features are a negative correlation over the central and eastern Pacific with a tail-like southeastward extension, and a more localized region of positive correlation over Indonesia. The negative correlation arises from the coincidence of anomalously higher heating rates during 1997–1998 with strongly negative values of the SOI (see http://www.bom.gov.au/climate/current/soi2.shtml).

[16] We note that a region of positive correlation occurs over the maritime continent, with a weak southeastward extension. One source is the coincidence of lower-than-average heating rates over the maritime continent with negative SOI values during 1997–1998. Another source of positive correlation is the occurrence of higher-than-average diurnal heating rates over northern Australia and the maritime continent during September 1988 and October 1998 (not shown), that coincided with positive values of the SOI.

[17] Time traces of different zonal wave number components corresponding to the anomalous January 1998 equatorial IR heating are indicated in Figure 5. The dominant component of the IR heating pattern is always the migrating (or W1) component. Because the phase of water vapor heating is zonally uniform in LT, westward and eastward traveling nonmigrating spectra are symmetric with respect to W1 (indicated in Figure 5 by the coincidence of solid and dashed colored curves). W1 is enhanced in 1997–1998, as a result of the greater longitudinal uniformity of heating over the tropical Pacific. During this same period, nonmigrating components exhibit a reduction in amplitude, relative to their time-averaged values. The implications of this heating spectrum for the tidal response will be explored further in section 4. Here, we note that the principal mode of the migrating diurnal tide has a deep vertical wavelength (~27 km), and propagates very efficiently into the mesosphere and lower thermosphere [Hagan, 1996]. Thus enhancement of the migrating component of the water vapor forcing during 1997–1998 may account in part for amplitude

Figure 3. (left) Latitude versus longitude of tropospheric mass-weighted average water vapor heating for (top left) January 1997, (middle left) 1998 and (bottom left) 1999. (right) Corresponding deviations from the climatological January heating.

Figure 4. Correlation between SOI and diurnal water vapor heating rate between computed from monthly values between 1988 and 2001.
enhancements observed in the middle atmosphere (Figures 1 and 2).

### 3.2. Latent Heating: Background

[18] Latent heat release by large-scale deep convection was first identified as a possible source of tidal excitation by Lindzen [1978] and Hamilton [1981]. Early estimates of diurnal latent heat release were inferred from rain gauge data, and a prescribed vertical structure for the latent heat release function. Although the quality of rain gauge data is very high, the network does not have the global coverage required to define tidal driving. One means of circumventing sampling deficiencies in space and in time has been to use diurnal heating rates from general circulation models.

**Figure 5.** Time dependence of the zonal wave numbers of solar heating at the equator, between 1995 and 2000. (top) Westward wave one (migrating) amplitudes. (middle) Nonmigrating amplitudes, as indicated in the legend. (bottom) Perturbation of zonal wave number amplitudes from 14-a climatological values. All values have been smoothed with a 3-month running mean.
as a proxy for measurement-based heating [Lieberman and Leovy, 1995; Miyahara et al., 1999].

[19] Our investigations of tidal driving due to latent release is an extension of the pioneering study of Forbes et al. [1997] (hereinafter referred to as F97). As described by F97, satellite-based global cloud imagery data from the Global Precipitation Climatology Project were used to derive latent heat estimates expressed as rainfall rates. Raw data were in the form of IR brightness temperatures; cold temperatures imply high cloud tops and deep convection in the tropics. Rainfall rates were estimated on $2.5^\circ \times 2.5^\circ$ longitude-latitude grid cells that contained a large number of pixels. F97 assumed that the rainfall rate scaled linearly with the fraction of pixels at temperatures below a threshold temperature. In other words, rainfall rate within a grid cell was assumed to be proportional to the area fraction which was "cold." The threshold temperature was taken by F97 as 235 K. The rational for this precise temperature threshold is not discussed by F97, but it is presumably based on experimentation and certainly low enough to exclude stratiform clouds. F97 then assumed that rainfall rate in mm h$^{-1}$ was equal to cloudiness fraction times 3. Diurnal and semidiurnal components of rainfall were determined by Fourier analysis of a composite year. The scaling between the diurnal component of cold cloud fraction and rainfall rate was justified post hoc, using extensive validation with ground-based rain gauges.

3.3. TRMM and ISCCP Data Analysis

[20] Two active missions are committed to the generation of global cloud and rainfall databases with sufficient resolution to derive the diurnal cycle: the International Satellite Cloud Climatology Project (ISCCP) and the Tropical Rainfall Measurement Mission (TRMM). ISCCP merges data from geostationary and polar-orbiting satellites in order to obtain infrared (IR) radiance data sets with high temporal and spatial resolution [Schiffer and Rossow, 1983]. The Tropical Rainfall Measuring Mission (TRMM) is a satellite mission jointly managed by NASA and the Japan Aerospace Exploration Agency (JAXA) that is designed to monitor and study tropical rainfall. The TRMM experiment commenced in 1997, and one of its goals is the provision of the four-dimensional distribution of rainfall and latent heating over oceanic and tropical continental regimes [Kummerow et al., 2000; Tao et al., 2006].

[21] Our initial efforts to quantify tidal driving due to latent heat release focus on the 3G68 data set, a public data set available from TRMM. This is a blended product making use of data from two different instruments aboard the TRMM satellite, the Precipitation Radar (PR) and the TRMM Microwave Imager (TMI). The 3G68 product includes hourly grids, composited from a month’s worth of data. Rainfall estimates derived for the PR, TMI and combined instruments are separately reported. In addition, stratiform and convective rainfall are reported for each grid point. For this study, convective rainfall estimates derived from the combined instruments were used to parameterize latent heating.

[22] The 3G68 data begin during January 1998 and have been continuously updated since that time. The files contain hourly values of rainfall composited over a month onto a fine spatial grid. The rainfall is partitioned into separate stratiform and convective fields. We use the convective rainfall estimates as a proxy for latent heat release. To evaluate tidal forcing, data are sorted into 3-h time bins, and are then spatially averaged into a coarse grid providing 10$^\circ$ resolution in longitude and 5$^\circ$ resolution in latitude.

[23] Examples of the resulting spatial maps of TRMM diurnal rainfall amplitude are shown in Figure 6 (left). TRMM amplitude maps are very similar in appearance to the results of F97. In contrast to diurnal water vapor heating (Figure 3), diurnal convection is concentrated much more strongly over tropical land masses than the oceans. The highest rainfall values occur over the Amazon basin, with
somewhat lower rates localized over central Africa. Diffuse secondary maxima are seen over India and the maritime continent. The corresponding longitudinal power spectrum (not shown) consists primarily of zonal wave numbers 5 and 3 traveling westward (W5) and eastward (E3), respectively. These patterns arise from the convolution of the diurnal solar heating function with a stationary zonal wave number 4 function that describes the equatorial land-sea contrast.

The phase of TRMM diurnal rainfall is shown in Figure 7 (left). Over tropical landmasses the hour of maximum is about 1500 LT. This afternoon convective rainfall maxima has been attributed to the afternoon static destabilization, and subsequent deep convective development due to solar heating [Byers and Braham, 1948; Blackadar, 1957; Wallace, 1975]. Over the oceans, the hour of maximum precipitation generally ranges between midnight and 0600 LT. The nocturnal maximum in precipitation over open ocean has been attributed to nocturnal long-wave cooling and subsequent static destabilization in cloud anvils [Kraus, 1963; Randall et al., 1991], a dynamical-radiative feedback mechanism whereby low level convergence and precipitation is favored at night [Gray and Jacobson, 1977], or nocturnal radiative moistening which acts to reduce entrainment and evaporation [Tao et al., 1996; Sui et al., 1997].

While TRMM products are state-of-the-art rainfall estimates, our focus on IA variability of the tidal forcing led us to explore other data sets for estimating tidal driving, due to TRMM’s relatively short period of operation. Mesospheric radar data from Kauai and other sites showed an anomalously large tide during 1997, at least twice the amplitude of a typical year. However, the TRMM data set commences in 1998, so it cannot be used to study the most interesting year. Moreover, we intend our study to be a comprehensive study of solar driving of the diurnal tide, including water vapor absorption as well as latent heating. The NVAP water vapor data set extends from 1988 to 2001, while TRMM extends from 1998 to the present, so there are only 3 a of overlap (1998–2001). These factors led us to make use of a global cloud imagery data set, in the spirit of the work done by F97.

ISCCP merges data from geostationary and polar-orbiting satellites in order to obtain infrared (IR) radiance data sets with full global coverage and high temporal and spatial resolution [Schiffer and Rossow, 1983]. ISCCP distributes several products from which global diurnal latent heat release can be inferred. Radiance (or brightness, or cloud top temperature) is used to infer cloud height, which is correlated with convective activity and rainfall [Arkin and Ardunay, 1989]. A number of studies provide cloud temperature thresholds for deep convective activity (DCA), typically at about 230K [Hendon and Woodberry, 1993; Williams and Avery, 1996; Forbes et al., 1997]. The DCA index is the excess of the observed over the threshold temperature, and is related to latent heat release via a rainfall rate parameter and the latent heat of condensation $L$. The vertical structure of latent heat release is not provided by ISCCP, but is commonly assumed to vary according to the vertical structure of tropical mesoscale convective systems [Frank and McBride, 1989; Houze, 1989; Mapes and Houze, 1992; Mapes and Houze, 1993].

Our latent heat calculations utilized the D1 archive (distributed by NASA Langley at http://eosweb.larc.nasa.gov/PRODOCS/isccp/table_sccp.html). The D1 data product provides the number of cloudy pixels on a global 280 km equal-area grid that provides 2.5° resolution at the equator. The temporal coverage extends from July 1983 through June 2005, with 3-h time resolution. The phase of cloudiness fraction at the $p_0 < 180$ hPa threshold showed the greatest qualitative and quantitative consistency with TRMM rainfall rate. We therefore produced monthly composites of 3-hourly cloudiness fraction at the $p_0 < 180$ hPa threshold.
threshold, on a $10^\circ \times 5^\circ$ longitude/latitude grid. When ISCCP cloud fraction and TRMM rainfall amplitudes were averaged over the continents (South America and Africa) and all months, cloudiness fraction at the $p_0 < 180$ hPa threshold was found to scale with TRMM diurnal convective rainfall by a factor of 1.8 mm h$^{-1}$.

[28] Comparisons in both amplitude and phase (hour of maximum) between diurnal TRMM rainfall rate and ISCCP scaled cloud fraction are shown in Figures 6 and 7. Continental-scale features are seen in both data sets. These include convection maxima over South America and Africa, India and the maritime continent, and sharp land-sea contrasts in phase. Generally, ISCCP patterns are smoother, whereas the TRMM patterns are smaller-scale.

[29] There are some further important differences between these two data sets. Compared to TRMM, ISCCP oceanic cloud fraction maximizes nearer to 0600 LT, whereas the TRMM often shows earlier (midnight) maritime convective maxima. The hour of maximum cloud fraction over the landmasses as seen in ISCCP occurs around 1700–1800 LT, whereas the phase of TRMM continental convection is generally closer to 1500 LT. This phase lag is interpreted as a byproduct of the typical life cycle of convective systems, which tend to produce most of their convective precipitation earlier in their life cycle than their maximum anvil extent [Houze, 1989]. The peak of convective precipitation, which TRMM infers from radar reflectivity measurements, is thought to be more closely spaced in time to the peak in actual latent heating than anvil extent, which is inferred from infrared brightness temperatures. Since this typical convective life cycle is phase locked to diurnal convective forcing over large areas of the tropics, this leads to the observed characteristic phase lag between convective precipitation and cloud anvil extent. This also serves to cause differences between infrared and other more direct estimates of precipitation [Liu et al., 2007].

[30] Our view of TRMM convective rate as the “ground truth” values led us to account for the TRMM-ISCCP phase discrepancies by retaining diurnal ISCCP cloudiness fraction amplitudes, but adjusting their phases on each pixel to coincide with the TRMM time-averaged phase at that grid point. This preserves the land-sea contrasts, but does not contain any seasonal variability in phase.

[31] To convert cloud fractions to latent heating rates, the vertical structure of the diurnal latent heating must be known. We adopt a vertical heating profile per unit mass $J(z)$ developed by Hong and Wang [1980]:

$$J(z) = A \left\{ \exp \left[ -\frac{z - 6.5}{5.39} \right]^2 - 0.23 \exp \left( \frac{-z}{1.31} \right) \right\}$$

We further assume that the surface rain rate $rr$ results entirely from the heating of the troposphere to altitude $z_t$ by $J$:

$$\int_0^{z_t} J(z) \rho(z) dz = Lrr$$

Assuming a constant temperature of 240K and a surface pressure of 1015 hPa in the evaluation of (2) yields a value for $A$ of 5.34 mW kg$^{-1}$ that corresponds to $rr = 1$ mm d$^{-1}$. Thus the conversion from ISCCP cloud fraction to latent heating consists of scaling cloud fractions by $1.8 \times 24$ to obtain $rr$ (in mm d$^{-1}$), and then by $J(z)$ with $A = 5.34$ mW kg$^{-1}$.

3.4. IA Variations of Diurnal ISCCP Cloud Fraction

[32] Monthly composites of the diurnal latent heating have been formed on the basis of ISCCP data between 1990 and 2005, hereafter denoted as climatology. Climatological heating patterns at the equator are generally centered over South America, Africa, and over or to the east of the maritime continent, as indicated in Figure 6. We have computed the variance of monthly ISCCP-derived estimates of latent heat release associated with periods between 12.9 months and 15 a. Figure 8 indicates that substantial IA variance occurs over South America, Africa, and the central western tropical Pacific. This pattern contrasts with the IA variance of diurnal water vapor IR heating (not shown), which is confined to the central eastern Pacific.

[33] For the present study, we confine our discussion of year-to-year latent heating variations to the patterns that prevailed during the 1997–1998 episode of anomalous water vapor heating. Changes in the tropical latent heating commenced in October 1997, and abated after April 1998. Figure 9 illustrates the heating for March 1998, which was characteristic of the October–April transition. The heating pattern stretches across nearly the entire tropical Pacific, as opposed to being contained in the western Pacific. The March 1998 heating maximum over the central eastern Pacific is approximately 5 mW kg$^{-1}$, compared with the climatological value of 0.6 mW kg$^{-1}$.

[34] Figure 10 shows the time traces of different zonal wave numbers for latent heat release at 2.5°S, a latitude chosen because it transects the anomalous pattern in the Pacific. In contrast to diurnal water vapor heating, latent heat release is described by a rich spectrum of nonmigrating wave components. The dominant wave numbers are generally the zonal mean ($S_0$), $E_3$, $W_2$ and $W_5$. In early 1998, several nonmigrating modes were enhanced, while $S_0$ remains clearly dominant. The perturbation amplitude spectrum (Figure 10, bottom) highlights the anomalous wave numbers structure during 1997–1998. $E_1$ and $W_3$ are enhanced, while $E_3$ (normally a leading mode) is diminished.

[35] The correlation between rainfall rate and SOI for 1990–2005 is shown in Figure 11. The pattern qualitatively resembles the correlation between water vapor heating and the SOI, shown in Figure 4. A broad center of negative correlation occurs over the central and eastern Pacific. The tail-shaped southeastward extension is also present, although situated further westward than in the solar-heating-SOI correlation pattern. The area of strongest positive SOI correlation is larger compared to the solar heating, extending over Indonesia and northern Australia. Overall the rainfall-SOI correlation pattern is “choppier” than the pattern in Figure 4, reflecting the more complicated and regional structures associated with deep convection.

4. Numerical Modeling

[36] The previous sections described tropospheric tidal sources, and identified a significant anomaly in the tropical water vapor heating during 1997–1998. The latent heat release functions also exhibited deviations from the climatological pattern during that same periods. These are the
Figure 8. IA variance of ISCCP diurnal latent heat release.

Figure 9. (left) Latitude versus longitude of tropospheric mass-weighted average diurnal latent heating release for (top left) March 1997, (middle left) 1998, and (bottom left) 1999. (right) Corresponding deviations from the climatological March heating.
same years when anomalously high mesospheric tidal amplitudes were observed at 5 tropical and subtropical observing facilities. This section explores quantitatively the effects of the heating anomalies upon the tidal response.

To compute the mesospheric tidal response, we use a time-dependent nonlinear primitive equation model described by Ortland and Alexander [2006]. The model incorporates the heating and cooling algorithm of Zhu et al. [1999]. Monthly climatologies of radiative gas concentrations input into the model include O$_2$, O, and CO$_2$ from the U.S. Naval Research Laboratory (NRL) Mass Spectrometer Incoherent Scatter Radar (MSIS) model [Hedin, 1996] and O$_3$ from the Upper Atmosphere Research Satellite (UARS) reference atmosphere [Swinbank and Ortland, 2003]. Radiative cooling from these species damps the tidal amplitude. Diffusive damping used in the model incorporates vertical profiles of molecular and eddy diffusion. We use the molecular diffusion values in units of m$^2$ s$^{-1}$ taken from Banks and Kocharts [1973]. The eddy diffusion is parameterized as by McLandress [2002a], and serves as a crude surrogate for the effects of gravity waves on the tide. Above 115 km ion drag is simulated with a Rayleigh friction rate of 15 d$^{-1}$. This basically serves as a sponge layer for the top of the model.

The tide is forced in the model by heating that varies sinusoidally at the diurnal harmonic with local time. The amplitude and phase of the oscillation at each latitude, longitude, and altitude grid point in the troposphere is determined from rainfall, water vapor or cloud data as described in section 3. A small diurnal response arises from the diurnal component of heating due to O$_3$ in the tropical stratosphere and mesosphere as determined by the Zhu et al. [1999] heating algorithm. The tide forcing is ramped up over 5 d and then maintained until a steady state is attained after 20 d of model integration.

The model is initialized with zonal mean zonal winds obtained from the COSPAR International Reference Atmosphere (CIRA) climatology of Fleming et al. [1990]. The temperatures are obtained from the background winds by using the thermal wind relation and the condition that the horizontal average temperature at a given pressure level equals the horizontal average of the CIRA temperature climatology. In order to prevent the model from relaxing to a state of radiative equilibrium, a constant forcing term that crudely represents the mean wind forcing due to gravity waves is added to the zonal momentum equation. The forcing is determined in off line zonally symmetric runs designed to maintain the model zonal mean winds at values close to the CIRA climatology. This method of forcing the zonal mean winds of the model was chosen because it does not affect the zonal mean component of the diurnal response.

We begin by showing the model temperature response to diurnal water vapor IR heating in January 1997–1999, in Figure 12. We focus initially on the temperature response because this field reflects the dominant latitudinal structure of the low-latitude tropospheric heating. Figure 12 is formatted identically to Figure 3, depicting the absolute and the “anomalous” temperature amplitudes arising from the corresponding water vapor heating. Examination of the anomalous responses to both water vapor and latent heating at tropical latitudes (not shown) reveals nearly uniform vertical structure. We therefore focus on results at 90 km, the model altitude closest to the level at which data are shown (in Figures 1 and 2).

At 90 km, January equatorial temperatures typically exhibit a minimum over the northern tip of South America, as evidenced by the 1997 plot (Figure 12, top left). In January 1998, the equatorial temperature structure is more zonally uniform, resulting in positive anomalies over South America and in the central Pacific Ocean. We note a

Figure 10. Time dependence of the zonal wave numbers of latent heat release at 2.5$^\circ$S, between 1995 and 2000. (top) Zonal wave number amplitudes, as indicated in the legend. Positive (negative) wave numbers denote westward (eastward) propagation. (bottom) Perturbation of zonal wave number amplitudes from climatological values. All values have been smoothed with a 3-month running mean.

Figure 11. Correlation between SOI and latent heat release rates, computed from monthly values of ISCCP cloud fraction between 1990 and 2005.
negative perturbation west of the maritime continent, that results from a northward shift of the climatological temperature pattern in 1998. The zonal wave number spectrum (not shown) is similar to that shown for the heating in Figure 5, indicating the dominance of the migrating tide.

Figure 12. (left) Latitude versus longitude of the amplitude of the model temperature response to water vapor heating at 90 km, for January (top left) 1997, (middle left) 1998, and (bottom left) 1999. (right) Corresponding deviations from the climatological January temperature amplitudes.

Figure 13. As in Figure 12, for March model temperature response to latent heat release.

[42] The 90 km level model temperatures response to latent heating is shown in Figure 13. This field is weaker and more structured in longitude than the water vapor response. Temperature amplitudes at this level in March typically show highly localized maxima across the Pacific.
Ocean and the Amazon, with a secondary maximum over the Indian Ocean. In 1998, the Amazon response is weaker, while the Indian Ocean response has shifted eastward to the maritime continent. We note that although the latent heat response is generally weaker than the water vapor response, the anomalies in their individual responses during 1998 are similar in magnitude. The zonal wave number spectrum (not shown) is similar to that shown for the heating in Figure 10.

The model response to the combined water vapor heating and latent heat release mechanisms is shown in Figure 14, for January and March 1998. The juxtaposition of the latent heating response results in “breaking up” of the nearly uniform longitudinal distribution of the water vapor heating response. The combined amplitude structure exhibits the highest amplitudes between 60°W and 120°E, and has a minimum over the central tropical Pacific.

The model meridional wind amplitudes corresponding to the temperatures in Figure 14 are shown in Figure 15. Amplitudes maximize near 20°, and nearly vanish at the equator, a pattern that reflects the dominance of the gravest

![Figure 14](image1.png)

![Figure 15](image2.png)
mode of the migrating diurnal tide [Chapman and Lindzen, 1970]. During March 1998 a pronounced hemispheric amplitude asymmetry is apparent, with stronger amplitudes in the southern hemisphere. The response to latent heat release introduces longitudinal variations into the amplitude, with the minima tending to occur in the central western Pacific. The largest deviation in wind amplitude occurs as a positive anomaly over South America and the Atlantic Ocean during January 1998.

Time series of modeled meridional wind amplitudes have been constructed at coordinates of MF observation sites. The analysis at Kauai coordinates is shown in Figure 16, the only location where the model behavior qualitatively mimics the MF data. The response to absorption of IR radiative heating dominates the model solution, with a secondary contribution from the latent heating response. The phase of the latent heating response lags the solar response by 15 h during boreal summer months, and about 7 h during the winter months. As a result, the net response tracks the solar heating response, but with a diminished amplitude due to the out-of-phase component of the latent heat response. Amplitude and phase have a pronounced annual cycle, with

Figure 16. Meridional winds at (23°N, 158°W), the model coordinates closest to Kauai. (top) Wind amplitude. (middle) Phase, as hour of maximum. (bottom) Deviation of amplitude from 1990 to 2002 monthly composite values. Solid red, model response to latent heat release. Solid blue, model response to water vapor heating. Solid black, sum of latent heating and water vapor response. Dotted lines, Kauai observations. All values have been smoothed with a 3-month running mean.
the amplitude of the model solar heating response peaking during the winter months. A positive amplitude anomaly in the net model amplitude is sustained through much of 1997, peaking in June of that year, disappearing by 1998, and reemerging in 1999. The source of both the 1997 and 1999 anomalies is a higher-than-average amplitude of the water vapor heating response.

Comparison of the observed (Kauai) winds to the modeled winds reveals some qualitative consistency. The observed phase shows an annual cycle that closely tracks that of the modeled phases. The observed amplitude was anomalously high in 1997 and much of 1999, as is the modeled amplitude. However, the model and the observations show substantial quantitative differences. Model amplitudes are more than twice the observed amplitudes. The phase of the observed winds lag the modeled winds by about 5–7 h. The anomalous amplitude behavior in the model (~5 m s\(^{-1}\) in 1997) is considerably weaker than the observed amplitude anomalies of 15 m s\(^{-1}\) in 1997 and 10 m s\(^{-1}\) in 1999. These differences may arise from model settings that do not reflect the actual atmospheric conditions between 1990 and 2002. Examples are the use of climatological mean winds and damping parameters, as opposed to actual, interannually varying values. Another possibility is the unrealistic modeling of tidal sources and sinks due to tide-gravity wave interaction [Orland and Alexander, 2006].

We close this section with a discussion of model wind evolution at two other locations. The purpose is to examine wind amplitude anomalies in the model MLT, and the various ways these can arise from water vapor heating and latent heating anomalies. Figure 17 illustrates two examples. During the later part of 1997, the amplitude of latent heat release response at (23°S, 56°W) (Figure 17, left) is very weak, and the solution is dominated by the response to water vapor heating. The amplitude of this response has a strong positive anomaly in late 1997, reflected in the net (or observable) amplitude. Whereas the model produced only a modest amplitude increase at Kauai in 1997, the amplitude increase at (23°S, 56°W) is ~8 m s\(^{-1}\), or ~15% of the

Figure 17. Meridional wind response at 90 km due to water vapor heating (blue curve), latent heat release (red curve) and their sum (solid black curve). (top) Amplitudes, (middle) phase (hour of maximum in UT), and (bottom) deviation from monthly composite amplitudes. (left) (23°S, 56°W), (right) (11°S, 158°W). All values have been smoothed with a 3-month running mean.
observed value. Thus our observation-driven model does produce significant deviations of tidal amplitudes in the MLT, albeit not at all the locations of available observations.

A different situation prevails at (11°S, 158°W). The latent heating response is significantly stronger than at 23°S, although still less than 50% of the water vapor heating response. However, the responses to water vapor heating and latent heating are out of phase. In late 1997, the amplitudes of both the individual responses showed positive deviations from their composite values, with the latent heating anomaly being larger. The amplitude of the net response, however, has the opposite tendency, being lower than its composite value. This paradoxical outcome results from the latent heating response being both higher-than-average and out-of-phase with the water vapor heating response. This results in the latent heating response “pulling down” the net response below its composite value. Thus it should be borne in mind that the sign of an observed amplitude anomaly may reflect the behavior of an individual dominant effect, or the interference of multiple effects.

5. Summary and Conclusions

This study focused on IA variations of tropospheric heating drives, and the MLT tidal response observed by radars and predicted by a model. The work was prompted by reports of tidal variability that have been linked to ENSO [Vial et al., 1994; Gurubaran et al., 2005], and by observations of MLT diurnal amplitudes enhancements at Hawaii and Christmas Island between late 1997 and early 1998.

We examined diurnal solar heating due to water vapor absorption, and diurnal latent heat release due to deep convection between 1988 and 2005. Water vapor heating is characterized primarily by a migrating component (W1). The climatological pattern of diurnal latent heat release is considerably more complex, projecting primarily onto E3, S0, W2 and W5.

Both water vapor and latent heating patterns exhibit anomalously higher amplitudes in the tropical central and eastern Pacific during 1997–1998. The weakening of a climatological “dry tongue” in the eastern tropical Pacific results in greater longitudinal uniformity of the tropospheric water vapor distribution. This structure produces a stronger W1 (migrating) component of water vapor heating. The patterns of latent heat release also show a greater Pacific extension between late 1997 and early 1998. This alteration is reflected by stronger projection into E1 and W3. We note that the perturbations in the amplitudes of the various modes (e.g., W1 associated with water vapor heating, and E1 and E3 associated with latent heat release) have similar orders of magnitude (see Figures 5 and 10). However, the magnitude of the (overwhelmingly migrating) water vapor heating drive is generally twice as strong as the latent heating drive.

Global patterns of diurnal water vapor and latent heating exhibit negative correlations with the SOI over the central and eastern Pacific. These patterns confirm earlier suggestions that year-to-year variability in diurnal forcing is linked to the ENSO phenomenon.

A primitive equation model was used to evaluate how water vapor heating and latent heat release contribute to diurnal winds in the mesosphere. The response to water vapor heating dominates the model solution, with secondary contributions from the latent heating response. Since the water vapor heating is characterized primarily by a migrating component, the migrating tide and its variability dominate the MLT wind. Nonmigrating components of the MLT tide are attributed primarily to the response to latent heat release. These elements can locally comprise up to about 50% of the diurnal solution.

Time series of model meridional wind amplitudes were compared to observations at Kauai, that showed a strong enhancement of tidal wind amplitude in 1997–1998. The net model response generally tracks the water vapor heating response, but with a diminished amplitude due to the out-of-phase contribution from a secondary latent heat response. A positive amplitude anomaly in the net model amplitude is sustained through much of 1997, disappearing by 1998, and reemerging in 1999. The source of both the 1997 and 1999 model anomalies at Kauai is a higher-than-average amplitude of the water vapor heating response.

We examined the model responses at other locations, where the magnitude and variability of the water vapor and latent heating responses “played out” differently than at Kauai. We have found that while both heating responses exhibit significant and comparable IA variability, the migrating component of the water vapor heating response is the main determinant of the net amplitude evolution. The reason is that the water vapor heating response generally dominates the latent heating response by a factor 2.

Although the timing of the modeled enhancements at Kauai is correct, the magnitude of the observed enhancements (~15 m s−1) is significantly larger than predicted by the model. The model does, however, simulate stronger enhancements at other latitudes and longitudes (~8 m s−1). We attribute the inability of the model to match the observations to differences between the model settings and the actual atmospheric conditions between 1990 and 2002.

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