Water's two height scales: The moist adiabat and the radiative troposphere

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SUMMARY

The temperature structure of the tropical troposphere resembles a moist adiabat, with a lapse-rate transition toward dry adiabatic where water becomes scarce at an altitude $H_{ma} \sim 8$ km (350 hPa). Infrared emission by water vapour cools a deeper layer, extending up to $H_{rd} \sim 14$ km (160 hPa). Five consequences of these unequal heights are reviewed.

1. Upper-tropospheric relative humidity is often low, highly variable, and bimodal, due to the rapidity of drying by radiative subsidence.

2. Large-scale divergent circulations (e.g. equatorial $\mathbf{u}$ wind) exhibit a two-celled vertical structure, with an elevated convergence layer near 8–10 km in the rising branch.

3. The dominant deep convective heating process changes from latent heating at low levels to eddy heat-flux convergence in the upper troposphere. This requires a substantial updraught-environment temperature difference, which leads to large entrainment near $H_{ma}$, yielding stratiform anvil clouds which also contribute radiative heating.

4. The rising branches of deep ($\sim H_{rd}$) vertical circulations export more heat than they import as moisture, so that large-scale tropical dynamics can be characterized by a 'gross moist stability'.

5. Divergent motions with a vertical wavelength $\sim 8$ km, corresponding to Kelvin or gravity wave speeds of $\sim 15$ m s$^{-1}$, are excited by simple (e.g. uniform) heating profiles extending through the lapse-rate change near $H_{ma}$.

KEYWORDS: Gross moist stability Radiative-convective equilibrium Tropical atmosphere Vertical structure

1. INTRODUCTION

The word troposphere comes from the Greek $\text{tropos}$, meaning 'turning'. Roughly speaking, the overturning layer of the tropical atmosphere is the layer which undergoes net radiative cooling, typically extending from the surface up to roughly 12–14 km† (see Fig. 2). Energy input to the atmosphere at the earth's surface, primarily in the form of water vapour flux, climatically balances this radiative cooling. The latent energy of water vapour heats the atmosphere where condensation occurs, while circulations redistribute this energy source to balance the radiative energy sink, yielding a radiative–convective equilibrium. To a good approximation, this bulk energy balance applies to the tropics alone, the exchanges with midlatitudes being much smaller (e.g. Fig. 13.12 of Peixoto and Oort 1992), so for the present purposes the tropics will be discussed in an idealized isolation.

The horizontal and temporal structure of the tropical general circulation are very complex, but certain aspects of its vertical structure are highly constrained by basic physical considerations. Specifically, stratification adjustment by gravity waves in the presence of moist convection (e.g. Bretherton 1993) keeps the mean temperature profile near a moist adiabat to a useful level of approximation. Meanwhile, water vapour steadily cools the troposphere by radiation, even where its concentration is quite low, owing to very active rotation bands and continuum effects (Doherty and Newell 1984; Clough et al. 1992). This paper explores consequences of these physical processes for the vertical structure of the troposphere and its circulations.

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† The tropical tropopause is often higher than this, due to dynamical effects that are beyond the scope of this paper (see, for example, Highwood and Hoskins (1998), Folkins et al. (1999) and Thuburn and Craig (2000)). Clouds strongly alter the radiative cooling over the clear-air profiles of Fig. 2 locally, but in the mean they do not change the sense and sign of this cooling (Bergman and Hendon 1998).
2. DATA AND METHODS

Mean thermodynamic soundings used in this study come from three tropical field programmes: the Coupled Ocean–Atmosphere Response Experiment (COARE, mean sounding from Zuidema (1998)) in the western near-equatorial Pacific; the Tropical Eastern Pacific Process Study (TEPPS, Yuter and Houze 1999); and the GARP (Global Atmospheric Research Program) Atlantic Tropical Experiment (GATE, the mean of the fast and slow line environmental soundings from Barnes and Siekman (1984)). Thermodynamic calculations were done using entropy formulae of Raymond and Blyth (1986) and Emanuel (1994), including contributions from precipitation and ice processes. In reversible ascent, condensate remains with the air, while in pseudo-adiabatic ascent the condensate is removed from the air as soon as it forms. All phase changes are assumed to occur at thermodynamic equilibrium, and relative humidity is expressed with respect to ice at temperatures below 0 °C. Equatorial diurnal-mean radiation for Fig. 2 was adequately computed using version 2 of the Community Climate Model (CCM2) radiation code, although it lacks the latest information on the water vapour continuum (Clough et al. 1992), which increases the cooling of the upper troposphere (Iacono et al. 2000), and hence strengthens the effects considered here. The spectral analysis of vertical modes in section 3(e) uses the Fulton and Schubert (1985) transform.

3. RESULTS

The mean stratification is first examined, followed by five subsections on physical applications.

Figure 1 shows mean temperature stratification profiles from two tropical field programmes (COARE and TEPPS). Also shown in Fig. 1 is the temperature of moist adiabats, both pseudo-adiabatic and reversible. The moist adiabatic lapse rates (defined as \(-dT/dz\), positive (where \(T\) is temperature and \(z\) is height)) are near 5 degC km\(^{-1}\) in the lower troposphere, then veer toward the dry adiabatic lapse rate of \(-10\) degC km\(^{-1}\) above ~8 km altitude \((H_{ma})\). The observed lapse rates are roughly similar, except in the boundary layer where they are nearly dry adiabatic, and above 12 km where they begin to decrease again, eventually becoming negative above the tropopause (>15 km). A large lapse rate is also seen in the 2–4 km layer, below the 0 °C level (Johnson et al. 1996), where a climatological cloudiness minimum prevails (Zuidema 1998).

A related quantity, important to thermally-driven circulations, is the static stability parameter \(\sigma = \partial / \partial p(s/C_p)\), where \(s\) is dry static energy and \(C_p\) is heat capacity at constant pressure \(p\). For moist adiabats, \(\sigma\) has a curved profile in the middle–upper troposphere, again matched roughly by the actual tropical stratification measurements between the boundary layer and about 12 km. The small-\(\sigma\) region in the upper troposphere, between about 8 and 13 km, has several interesting consequences, as illustrated in the following five subsections.

(a) Radiative drydown

Water vapour is the dominant radiatively active gas even in dry regions of the troposphere (Harries 1997). Upper-tropospheric humidity is especially important to the earth’s radiation budget, but is subject to large observational and model uncertainties. Brown and Zhang (1997) showed that upper-tropospheric humidity distributions in COARE soundings were bimodal, with a dry mode and a wet mode and relatively few soundings sampling the mean condition in between. This behaviour is partly a consequence of the small-\(\sigma\) layer in the upper troposphere.
In the tropical free troposphere, temperature is nearly constant in the horizontal and in time, with clear-air radiative cooling balanced by vertical entropy advection in the subsiding branch of vertical circulations. Neglecting temperature change and horizontal advection, the thermodynamic equation reduces to $\omega \sigma \sim Q$, where $\omega$ is pressure vertical velocity and $Q$ is heating rate, expressed as a temperature tendency. This defines the diabatic vertical velocity $\omega_d = Q/\sigma$.

Consider a cloudless column with the COARE mean temperature profile, initially at 85% relative humidity. It undergoes a net radiative cooling that is initially $\sim 1.5$ degC per day below the 12 km level, as shown by the full line in Fig. 2(b). Above 1 km, assume that this cooling is balanced by diabatic subsidence, $\omega_d > 0$, and that the specific-humidity ($q$) equation has no horizontal advection or source terms, so that

$$\frac{\partial q}{\partial t} + \omega_d \frac{\partial q}{\partial p} = 0.$$  \hspace{1cm} (1)

The evolution of relative humidity according to (1) over a 5-day period is shown in Fig. 2(a), while the corresponding evolution of the radiative cooling profile is in Fig. 2(b).

Figure 2 indicates that the upper troposphere dries out very rapidly: at 11 km (250 hPa), humidity decreases to less than half its initial value in just 1 day. This rapid drying follows in part from the small value of $\sigma$ in the upper troposphere, which requires large subsidence rates to balance radiative cooling (see also Eq. (3) and Fig. 3 of Sherwood 1999). The drying of the upper troposphere enhances the cooling rate of the middle troposphere so that the radiative–subsidence drying works its way downward.
with time. The bimodal distribution of upper-tropospheric humidity can be understood as a consequence of the rapidity of this drying: after being moistened by transports from below, upper-tropospheric air spends very little time with moderate values of relative humidity (Salathé and Hartmann 1997).

\[ (b) \text{ Convergence at } \sim 10 \text{ km in convective regions} \]

Consider a two-box model of tropical vertical circulations: a dry box containing only radiatively driven subsidence, as in subsection 3(a), and a rainy box containing a corresponding net heating (an excess of convective heating over radiative cooling) and hence a corresponding mean upward motion ($\omega_d < 0$). The mass continuity equation requires a resulting horizontal wind between the boxes, the vertical profile of which is given by:

\[ \delta_d = - \left( \frac{\partial \omega_d}{\partial p} \right) = - \frac{\partial}{\partial p} \left( \frac{Q}{\sigma} \right) \]

which defines the diabatic divergence $\delta_d$ (Mapes and Houze 1995). Features of the profile of $\sigma$ are enhanced by this differentiation, so that $\delta_d$ may actually change sign with height, even for a smooth, single-signed heating profile.

Figure 3 illustrates such a two-box model, subject to the smooth equal and opposite heating and cooling profiles shown on the left and right sides of Fig. 3. The corresponding profile of $\omega_d$ is overlaid (with sign reversed) in the heated box. Profiles of $\delta_d$,
Figure 3. The vertical structure of diabatic divergence $\delta_d$ (interbox wind, panel (b)) between heated and cooled regions with the heating profiles, $Q$, plotted in panels (a) and (c). The cooling profile is idealized from a time average from Fig. 2(b), but is exactly uniform with height so that wind structure below 10 km arises solely from the profile of the static stability parameter $\sigma$. Below the 850 hPa level, where evaluating Eq. (2) is problematic, a constant mass-conserving divergent wind is assumed to close the circulation. The diabatic vertical velocity, $-\omega_d$, profile is shown by a dotted line in (c).

and hence of the divergent (interbox) wind necessary to maintain the assumed constant temperature and $\sigma$ in the face of this differential heating, are plotted in the middle of Fig. 3, for three different $\sigma$ profiles shown in Fig. 1: COARE (dash), TEPPS (short dash), and the reversible moist adiabat (full). In addition to the low-level flow toward the heated box, and upper-level flow toward the cooled box, there is a convergence into the heated box near 9–10 km (300 hPa) for both the observed and moist-adiabatic $\sigma$ profiles, due to the small-$\sigma$ region in the upper troposphere. The profiles using observed $\sigma$ also show enhanced mass flows out of the heated box near 1.5 and 4.5 km, with an inflow layer in between.

Many divergent wind profiles in the tropics resemble this two-box model profile of $\delta_d$. Perhaps the grandest example is the zonal-mean meridional flow on the equator between winter and summer hemispheres. Figure 4 shows the May–August zonal-mean meridional wind at the equator, as depicted in the NCEP* and ECMWF† reanalyses. A two-celled circulation is evident, with southerly flow into the northern hemisphere at 400–500 hPa in addition to the main near-surface inflow, and slight northerly flow at 600–700 hPa in addition to the main outflow at 150–200 hPa. Similar structure can be seen in the meridional wind at individual longitudes, from Christmas Island in the central Pacific (K. Gage, personal communication), to COARE soundings in the western Pacific (Lin and Johnson 1996), to soundings in the Bay of Bengal during the Asian

* National Centers for Environmental Prediction.
† European Centre for Medium-Range Weather Forecasts.
summer monsoon (JASMINE, Y. Serra, personal communication). Divergence analyses in various regions also show similar structure (N. Nishi, personal communication).

The tendency for a convergence layer near 8–10 km (or ~400 hPa) in convecting regions is present on smaller scales as well, e.g. in divergence profiles from GATE (Thompson et al. 1979), the Australian monsoon (Mapes and Houze 1993), and the COARE Intensive Flux Array (Lin and Johnson 1996). This convergence layer was interpreted as the break between the divergent outflow layers associated with two distinct convective cloud populations by Johnson et al. (1999), whose Fig. 13 also highlights its relationship to the small-stability layer of the upper troposphere. The observation that the 400 hPa convergence layer is diurnally varying (e.g. Fig. 5 of Albright et al. 1981) remains difficult to interpret, since convective and radiative heating both vary diurnally, in both their intensities and profiles.

The uniform smooth heating profiles and assumptions behind Fig. 3 are not realistic in detail, but the point here is that these secondary divergent wind-profile features may be expected to emerge irrespective of special heating-profile shapes.

(c) Eddy heating above $H_{ma}$: a reason for CAPE and stratiform precipitation

In radiative–convective equilibrium, the radiative cooling of the upper troposphere must be balanced by deep convecting heating. However, latent-heat release along a moist adiabat decreases rapidly above $H_{ma} \sim$ 8 km, as water amounts become small. Deep convection heats the upper troposphere largely by eddy heat-flux convergence, such that latent heat released in the lower troposphere is redistributed upward by warm updraughts (Houze 1982; Sui et al. 1994; Grabowski et al. 2000). This eddy heating by convection depends on the upper troposphere being significantly colder than the updraughts. For simplicity, consider the simplest idealization for updraught temperature: a parcel ascending along a moist adiabat.

Figure 5 shows the temperature excess of a parcel rising from low levels in the COARE mean sounding (TEPPS and GATE results are similar). To bracket microphysical uncertainties, both pseudo-adiabatic and reversible processes are indicated. No mixing is considered. At the 9–10 km level, the pseudo-adiabatic parcel is 4 degC warmer.
than the environment, and the reversible parcel nearly 8 degC. The greater warmth of the reversible parcel comes both from the latent heat of freezing of the suspended condensate, and from the thermal inertia of the condensate, which delays the parcel's cooling with height.

A large updraught-environment temperature difference tends to correspond to large updraught buoyancy, and hence convective available potential energy (CAPE, buoyancy integrated over height). To show this, Fig. 5 also indicates parcel buoyancy, expressed as a parcel-environment difference in density temperature $T_p$, which includes water effects on density. The upper-tropospheric buoyancy is similar for pseudo-adiabatic and reversible ascent (as discussed by Williams and Renno (1993)), and is substantial, with about half (pseudo-adiabatic) to $\frac{2}{3}$ (reversible) of CAPE accounted for by parcel buoyancy above $H_{ma}$ (8 km). The existence of a reservoir of CAPE might therefore be viewed as another consequence of the mismatch between $H_{ma}$ and $H_{rad}$. The relationship between this explanation of CAPE and explanations based on entropy considerations (Renno and Ingersoll 1996; Craig 1996; Emanuel and Bister 1996) remains unclear.

In any case, the CAPE of unmixed parcels in averaged temperature and moisture soundings is not very relevant to real convective processes. There are two ways an increase in eddy heating above $H_{ma}$ can be accomplished: through increases of the temperature excess of updraughts, or increases of the updraught mass flux (entrainment). Since entrainment is driven by vertical buoyancy gradients (Taylor and Baker 1991), the former tends to cause the latter, so that the increase is likely to be split between these mechanisms. In the light of section 3(b)'s result that radiative cooling of clear areas tends to drive a convergent flow into cloudy areas, it seems inevitable that convection should entrain mass near the 8–10 km level.
Distinctive upper-tropospheric maxima in vertical velocity have been observed in COARE convective cells, both observed and modelled (Trier et al. 1997). At the same time, the eddy heat fluxes by mesoscale updraughts in stratiform precipitation areas associated with deep convection are substantial (Houze 1982). For microphysical reasons, a somewhat discrete stratiform process is often portrayed, with the detrainment of convective cell mass fluxes developing a mesoscale cloud mass in the middle–upper troposphere, which then ascends hydrostatically for several hours after the convective cells have decayed (Houze 1997). But these convectively-driven mesoscale updraughts are functionally equivalent to enhanced middle–upper tropospheric entrainment into convection. For example, Robe and Emanuel (1996) and Tompkins and Craig (1998) showed that mass flux, convective draught area, and cloudy area increase suddenly with height above 8 km in statistical–equilibrium convection, in periodic cloud-resolving model domains too small to contain mesoscale cloud clusters. Locally, upper-tropospheric cloud also undergoes radiative heating, although not enough to cancel tropical upper-tropospheric cooling on a global basis.

In summary, it appears that stratiform anvils are spawned by deep convection as a result of its need to heat the upper troposphere, between \( H_{ma} \) and \( H_{rad} \). In a climatic sense, radiation lowers the temperature of the upper troposphere until a balance is achieved with convective and mesoscale eddy heating and cloud radiative heating, all three of which may be ultimately traced to the large temperature excess of convective updraughts in this layer.

\[ \text{(d) Gross moist stability} \]

The previous subsection discussed why the middle–upper tropospheric temperature must be less than that of updraughts, and therefore (approximately) of a moist adiabat representing boundary-layer conditions. It follows that there must be a minimum of moist static energy \( h (h = C_p T \text{ (thermal energy)} + gz \text{ (potential energy)} + Lq \text{ (latent energy)}) \) somewhere in the troposphere. Although the temperature difference is maximum at \( \sim 10 \text{ km} \) (Fig. 5), the \( h \) minimum is lower because of the strong contribution of latent heat, \( Lq \), which is confined below \( H_{ma} \sim 8 \text{ km} \). The exact shape of the \( h \) profile depends on the profile of relative humidity (RH), which is highly variable and beyond the scope of this paper. Nonetheless, the layer of low \( h \) is highly confined to the lower troposphere for a wide variety of RH profiles. For example, the left side of Fig. 6 shows moist static energy profiles for the COARE, TEPPS, and GATE mean soundings, along with profiles for COARE temperatures with 60 and 80% relative humidity.

Since \( h \) is conserved with condensation, the column-integrated \( h \) budget has only radiation and surface fluxes as sources. Riehl and Malkus (1958) showed that, for the earth’s equatorial trough zone, the column integral of horizontal \( h \) flux divergence is positive, meaning that more \( h \) is exported as dry static energy in the upper troposphere than is imported as moisture at low levels. This \( h \) export, along with net radiative cooling, is balanced by surface fluxes. In transient disturbances, the net export of \( h \) from convecting areas has been hypothesized to cause local \( h \) decreases (cooling and drying), which act as a restoring force on vertical displacement. As a result, the tropical troposphere presents a reduced stratification or ‘gross moist stability’, so that large-scale waves propagate, but at a slower speed than they would have in a dry atmosphere (Neelin and Held 1987; Emanuel et al. 1994; Yu et al. 1998).

The reason for net horizontal divergence (export) of \( h \) from regions of upward motion is that the low-\( h \) layer is concentrated in the convergent lower half of circulations which span the whole depth of the troposphere. For example, Fig. 6(b) shows the divergence profile characterizing the lowest mode excited by tropospheric heating (see
next subsection for details). Its mass-weighted mean is zero when integrated to 15.7 km, but the mass-weighted integral of its product with $h$ is positive for all the $h$ profiles shown, indicating gross moist stability for a wave with this vertical structure. This will be true in general for any deep simple divergence profile spanning the whole troposphere, such that convergence predominates through the lowest several kilometres. However, this gross moist stability integral is almost exactly zero for the hemispheric divergence ($v$) profiles of Fig. 4, and for the similarly-shaped dashed divergence profile in Fig. 6.

Gross moist stability provides a simple theory for travelling large-scale disturbances in a moist convecting atmosphere, but is incomplete. Even for disturbances with deep convergence structure, the balance may be tipped to ‘gross moist instability’, meaning that regions of upward motion are able to persist and grow without propagating, by cloud-radiative effects (Raymond 2000; Lee et al. 2001) or by condensation-driven surface flux enhancements, as in tropical cyclones.

As a sidelight, comparison of Figs. 5 and 6 cautions that the common practice of assessing parcel buoyancy as the difference between parcel $h$ (or equivalent potential temperature $\theta_e$) and saturation $h$ of the environment gives a quite distorted (although correct in sign) view of the buoyancy profile, owing to the nonlinear dependence of $h$ on temperature at low levels (through the $Lq$ term).

(e) Vertical modes of the atmosphere excited by heating and cooling

The previous subsections have concentrated increasingly on the small difference between observed tropical temperature profiles and moist adiabats. In this final subsection we return to the quite usefully accurate approximation (for dynamical wave propagation) that tropical temperature stratification may be regarded as nearly moist adiabatic.

Figure 7 shows profiles of the Brunt–Väisälä frequency $N = \left\{ \frac{g}{\theta} (\partial \theta / \partial z) \right\}^{1/2}$ for the tropical soundings and moist adiabats. The moist adiabats swerve from values above
0.01 s$^{-1}$ below 8 km to smaller values in the upper troposphere. The observed profiles are very similar to each other, and differ from the adiabats in having low-$N$ boundary layers near the surface and high-$N$ above $\sim$12 km. Interesting features are also seen in the lower troposphere, below $\sim$5 km. These features are related to climatological cloudiness layers (Zuidema 1998). Might they be ultimately related to ice processes near the 0 °C level (4 km) (Johnson et al. 1996)?

The effect of this Brunt–Väisälä profile on dynamical waves can be addressed using the Fulton and Schubert (1985) vertical transform, which solves the vertical structure equation derived from linearized primitive equations for arbitrary stratification profiles. The use of this transform for solutions to the dynamical response to heating is discussed by Fulton and Schubert (1985), Mapes and Houze (1995) and Mapes (1999). For present purposes, it suffices to say that the transform returns an amplitude spectrum characterizing the amplitudes, vertical structures and associated gravity-wave phase speeds $c$ excited by a given heating profile. On a rotating sphere, heating excites a family of waves, not just gravity waves, but the family can still be characterized by $c$, or alternately by the ‘equivalent depth’ $H = c^2/g$.

Consider the simple heating/cooling profile depicted by full lines in Fig. 3 (constant below the 325 hPa level, zero above the 175 hPa level). Figure 8 shows spectra of the vertical modes excited by such a heating or cooling process, operating within three different atmospheric stratification profiles. The spectra are expressed in terms of the amplitude of the projection of the diabatic divergence profile, Eq. (2), onto each vertical mode, as a function of the mode’s value of $c$. Figure 8 indicates that this heating profile excites two main fast spectral bands*, near 50 m s$^{-1}$ and 15 m s$^{-1}$, in either the COARE or moist-adiabatic stratifications. In contrast, this same heating excites a much broader spectrum of phase speeds in an isothermal (constant-$N$) atmosphere.

*The spectra are discretized by an upper rigid-lid boundary condition, here at 0.1 hPa. With the upper lid at other altitudes, different discretizations of the spectrum occur, but the bands are robust (Mapes 1999).
These two dominant vertical wavelength bands are thus consequences not of the heating-profile shape, but rather of the low-stability layer in the upper troposphere in any stratification resembling a moist adiabat. The vertical structures of the two bands are indicated in Fig. 6, which shows two spectral truncations of the wind profile of Fig. 3 (without the need for Fig. 3's artifice in the lowest 1.5 km). The ~50 m s$^{-1}$ band is a half-wave through the troposphere (full line in Fig. 6(b)). The 15 m s$^{-1}$ band is a wavelength related to the difference between $H_{ma}$ and $H_{rad}$, and expresses the convergence profile feature near 8 km (400 hPa) discussed in section 2(b). But because the spectral band is narrow, it is wavelike in the vertical and hence includes also the sharpness of divergence near the tropopause, divergence near 600 hPa, and sharpened convergence near the surface (dashed line in Fig. 6(b)).

The two-band truncation of this response to a uniform heating profile (Fig. 6, dashed line) bears considerable resemblance to the observed profile of equatorial $v$ (Fig. 4), with wave numbers $\frac{1}{2}$ and 2 spanning the tropospheric layer. The divergence at 600 hPa in Fig. 4 appears to have a counterpart in Figs. 3 and hence 6, in the dashed and dotted lines arising from the somewhat mysterious deviation of observed stratifications from moist adiabats. Might this deviation be a spectral ringing phenomenon emanating from the upper-tropospheric low stability layer above $H_{ma}$? This question refers to the distinctive 'melting mode' (with $c \sim 6$ m s$^{-1}$) inside mesoscale convective systems noticed by Mapes and Houze (1995), in which a localized forcing at a fixed altitude (melting) apparently develops into a wavelike pattern in the vertical. The hypothesized mechanism involved internal waves, which are very dispersive with respect to vertical wave number,
interacting with convection. Or is this deviation a local feature driven directly by melting (Johnson et al. 1996)? More insight into these questions might be found in the time-mean stratification profiles of cloud-resolving models with and without ice processes.

Projection spectra like Fig. 8 of wind and geopotential height variations measured by COARE radiosondes were examined by Mapes (1999). Variability was observed in three bands, near 50, 25, and 15 m s\(^{-1}\). At that time, explanation was lacking for the 15 m s\(^{-1}\) band (that nagging lack was one of the motivations for this study). In the present discussion, the 25 m s\(^{-1}\) band has not appeared, based only on deep simple heating profiles and nearly moist adiabatic stratifications.

If the heating profile has more structure within the troposphere, that can preferentially excite other vertical wavelengths. For example, the light full line in Fig. 8 shows the projection spectrum excited by a half-troposphere heating profile (identical to the heating profiles of Fig. 3 but with the top shifted down 200 hPa, i.e. constant heating below the 550 hPa level). A \(\sim 23\) m s\(^{-1}\) spectral band (corresponding to wave number 1 of the troposphere) is strongly excited. This mode is also strongly excited by the stratiform rain regions of tropical mesoscale convective systems (Mapes and Houze 1995; Houze 1997).

The lower-tropospheric cooling in stratiform rain regions is apparently balanced in the climatic sense by lower-tropospheric condensation heating by a population of cumulus congestus showers reaching the middle troposphere (Johnson et al. 1999). In other words, activity in vertical wave number 1 of the troposphere is apparently dependent on the interplay between cancelling aspects of two different populations of moist convective cloud systems (Mapes 2000). Since both these populations are parametrized phenomena in large-scale models, the dynamics of tropospheric wave number 1 are highly dependent on model details, or even on vertical resolution with a given model (Innes et al. 2001). Perhaps the bewildering diversity of global model simulations of tropical convective variability is due in part to haphazard and undiagnosed differences in the excitation of this active (in observations) but parametrization-dependent (in models) vertical mode.

4. Summary and Discussion

There is a mismatch between the \(H_{\text{ma}} \sim 8\) km layer in which moist adiabatic ascent releases substantial amounts of latent heat, and the \(H_{\text{rad}} \sim 12\) km layer in which radiative cooling by water is strong. The result is a climatological layer of small static stability in the upper troposphere. The combination of this stratification profile with diabatic processes, which ultimately take their vertical scales from the radiative cooling profile, has several interesting consequences. Most of these are previously known, but have been synthesized here.

One important aspect of vertical structure in the tropical troposphere apparently does not flow directly from moist adiabatic and radiative processes: wave-number 1 structure through the depth of the troposphere. This wavelength is excited by cumulus congestus showers (which heat the lower half of the troposphere), and stratiform precipitation in organized mesoscale deep convective systems (which cools it). These processes cancel each other on average, but gradients in their geographical and time distributions drive circulations in this mode, which may be a controlling factor for some forms of large-scale convective variability (Mapes 2000).

Questions linger about deviations of the observed stratification from moist adiabatic, for example the small-\(\sigma\) (Fig. 1), low-\(N\) (Fig. 7) layer between 2–4 km observed in all three mean soundings examined here. This feature is apparently linked to the minimum of cloudiness in this layer (Zuidema 1998), as the detrainment of air from convective
clouds is reduced by the low stability (Johnson et al. 1999). Might this structure be linked to melting processes at the 0 °C (~5 km) level (Johnson et al. 1996), or might it be a spectral ringing phenomenon (section 3(e))? Credible (e.g. cloud-resolving) radiative-convective models with and without ice processes might provide answers.

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