Modeling Tropical Convergence Based on the Moist Static Energy Budget

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ABSTRACT

The vertically integrated moist static energy equation provides a convenient starting point for the construction of simple models of the time-mean low level convergence in the tropics. A vertically integrated measure of the moist static stability, the "gross moist stability," proves to be of central importance. Minima in this quantity mark the positions of the tropical convergence zones. We argue that the positions of these minima are determined by the time-mean moisture field, which is, in turn, closely tied to the time-mean surface temperature.

1. Introduction

The problem of relating tropical heating of the atmosphere to sea surface temperature (SST) has recently received new prominence in connection with modeling of the El Niño–Southern Oscillation (ENSO) phenomenon. An important role has been played in ENSO studies by idealized atmospheric models which attempt to provide the time-mean surface stress distribution in the tropics as a function of the SST field. Linear dissipative models for the low level flow have proven to be of considerable value when forced by a given latent heating distribution (Matsuno, 1966; Webster, 1972; Gill, 1980). Our own work in mimicking a GCM's behavior with idealized models confirms the result that a useful time-mean low level flow can be obtained with simple linear models, given the latent heating, or (what is almost equivalent) that the rotational part of the low level flow can be obtained, given the divergent part. To employ such models for ENSO studies, the atmospheric heating or low level divergence must be obtained as a function of the oceanic boundary conditions.

It has long been observed that regions of high precipitation are correlated with regions of high sea surface temperature (e.g., Bjerknes et al., 1969; Manabe et al., 1974; Cornejo-Garrido and Stone, 1977; Liebmann and Hartmann, 1982; Gill and Rasmussen, 1983), although the reason for this correlation has not been agreed upon. Based on this observation, it has been common to parameterize heating anomalies upon SST anomalies in one fashion or another (e.g., Webster, 1981; Zebiak, 1982; Gill, 1985; Philander et al., 1984; Anderson and McCreary, 1985). However, it has also been observed that SST anomalies in a region of high climatological SST (i.e., the western equatorial Pacific) have a stronger influence on the heating field than do comparable SST anomalies in colder regions (i.e., the eastern equatorial Pacific). In other words, SST influences the heating field in a nonlinear manner, with the result that heating anomalies can be considerably offset from SST anomalies (Gill and Rasmusson, 1983; Rasmusson, 1984). Indeed, it is sometimes stated that deep convective heating becomes especially intense when total SST approaches a critical temperature ($\approx 28°$; Gadgil et al., 1984). This temperature dependence cannot simply be a result of the nonlinearity of the Clausius-Clapeyron equation making itself felt through evaporation; the nonlinearity is much too weak. Furthermore, in regions of intense time-mean latent heating in the tropics, precipitation is primarily balanced by low level moisture convergence rather than evaporation (e.g., Cornejo-Garrido and Stone, 1977; Shukla and Wallace, 1983; Khalsa, 1983). As a result, tropical precipitation is not well correlated with evaporation either for climatology or ENSO anomalies (Baumgartner and Riechel, 1975; Khalsa, 1983). The low level flow carrying the moisture is, in turn, driven by the heating. Given this feedback, the link between SST and heating seems complex. Some attempts have been made to include a parameterized "convergence feedback" in which a fixed initial heating is amplified by latent heating (Webster, 1981; Zebiak, 1986). Other current attempts at constructing idealized atmospheric models for ENSO studies are directed toward inclusion of a simple moisture equation, but with basically the same process of amplifying a sensible heating anomaly (e.g., Gill, 1985).

One might ask why a convective parameterization scheme such as Kuo's (1965) or Arakawa and Schubert's (1974), coupled with an appropriate boundary layer theory, could not be added to the equations for the large scale flow and moisture so as to produce a model for the time-mean tropical heating and circulation. The difficulty is that these convective parameterizations induce nontrivial large scale transients (through CISK in particular). It is not simply a convective parameterization as usually formulated, but a finite amplitude CISK theory that would be required to model the time-mean heating.

In light of this complexity, it is appropriate to look for constraints that guide one towards a plausible simple model, or for ways in which uncertainties can be lumped into a small number of parameters. We argue here that use of the vertically integrated moist static energy budget can give considerable insight into the maintenance of the time-mean tropical convergence and precipitation patterns, and that a quantity we refer to as the "gross moist stability" provides a convenient way of summarizing our ignorance of the details of the convective and large scale transients. Plausible, but ultimately ad hoc, assumptions for the behavior of the gross moist stability lead to simple models of the mean low level convergence and precipitation patterns in the tropics.

The constraints imposed on the flow by moist static energy considerations are outlined in section 2. In section 3, we present a simple model of the low level convergence field as a function of the surface temperature, based on particular assumptions regarding the behavior of the moist static stability. In section 4, we show that the simplest version of the model can account qualitatively for the climatological seasonal movements of the convergence zones and for their movements in El Niño episodes. Possible criticisms of our approach are discussed in section 5.

2. The moist static energy formulation

The dry and moist static energies are defined as

$$s = C_pT + \Phi; \quad m = s + Lq \tag{2.1}$$

where Φ is the geopotential, L the latent heat of condensation and q the specific humidity. The dry static energy equation may be written in pressure coordinates as

$$\partial_t s + \nabla \cdot s\mathbf{V} + \partial_p s\omega = g\partial_p(F^{\mathrm{R}} + F^{\mathrm{S}}) + Q_{\mathrm{LH}} \tag{2.2}$$

where F^{R} and F^{S} are respectively the upward directed vertical fluxes of energy due to radiation and diffusion of sensible heat, and Q_{LH} is the latent heating per unit mass. The diffusive flux, F^{S}, is included for the sake of the lower boundary where the vertical motion is zero but the flux of sensible heat is not. The main approximation in the derivation of (2.2) is the neglect of kinetic energy transport, which is an excellent approximation for our purposes. To derive the moist static energy equation, one employs the moisture equation

$$\partial_t Lq + \nabla \cdot Lq\mathbf{V} + \partial_p \omega Lq = -Q_{\mathrm{LH}} + g\partial_p F^{\mathrm{L}} \tag{2.3}$$

where F^{L} is the vertical latent heat flux due to diffusion of moisture. Substituting for Q_{LH} in (2.2) yields

$$\partial_t m + \nabla \cdot m\mathbf{V} + \partial_p m\omega = g\partial_p F \tag{2.4}$$

where $F = F^{\mathrm{R}} + F^{\mathrm{S}} + F^{\mathrm{L}}$ is the total energy flux. At the surface, $F^{\mathrm{L}} = LE$, where E is evaporation. At the top of the atmosphere, $F = F^{\mathrm{R}}$.

Assuming that $\omega = 0$ at top ($p = p_{\mathrm{T}}$) and bottom ($p = p_{\mathrm{B}}$) boundaries, and using the notation

$$\langle A \rangle = \int_{p_{\mathrm{T}}}^{p_{\mathrm{B}}} A \frac{dp}{g},$$

we have

$$\langle \partial_t m \rangle + \langle \nabla \cdot m\mathbf{V} \rangle = F_{\mathrm{B}} - F_{\mathrm{T}} \tag{2.5}$$

where subscripts T and B denote top and bottom. Motivated by the observation that the large scale divergence in regions of intense convection tends to have a simple vertical structure in the tropics, with one sign near the ground, opposite sign near the tropopause, and small values in the middle troposphere, we also introduce the midtropospheric level p_{M} and define upper and lower layer quantities

$$\nabla \cdot \mathbf{V}_2 \equiv \int_{p_{\mathrm{M}}}^{p_{\mathrm{B}}} \nabla \cdot \bar{\mathbf{V}} \frac{dp}{g} = -\nabla \cdot \mathbf{V}_1$$

$$m_2 \equiv \int_{p_{\mathrm{M}}}^{p_{\mathrm{B}}} \bar{m} \nabla \cdot \bar{\mathbf{V}} \frac{dp}{g} (\nabla \cdot \mathbf{V}_2)^{-1}$$

$$m_1 \equiv \int_{p_{\mathrm{T}}}^{p_{\mathrm{m}}} \bar{m} \nabla \cdot \bar{\mathbf{V}} \frac{dp}{g} (\nabla \cdot \mathbf{V}_1)^{-1}$$

$$\Delta m \equiv m_1 - m_2. \tag{2.6}$$

Denoting time averages as $\overline{(\ )}$, departures from time averages as $(\ )'$ and neglecting the averages of time derivatives, we rewrite (2.5) as

$$-\Delta m \nabla \cdot \mathbf{V}_2 + \langle \bar{\mathbf{V}} \cdot \nabla \bar{m} \rangle + \langle \nabla \cdot \overline{m'V'} \rangle = \bar{F}_{\mathrm{B}} - \bar{F}_{\mathrm{T}}. \tag{2.7}$$

The normalized divergence acts as a vertical weighting of m in the integrals of (2.6). However, the vertical structure of the divergence may be complex in regions removed from the convection, and thinking in terms of a two layer model may not always be useful. We return to this difficulty in section 5.

We define dry static energy and moisture variables s_1, s_2, q_1 and q_2 by exact analogy with (2.6), and set $\Delta s \equiv s_1 - s_2$ and $\Delta q \equiv q_2 - q_1$, so that $\Delta m = \Delta s - L\Delta q$. Here Δq has been defined with the opposite sign convention so as to be positive. With the same manipulations that lead to (2.7), we also have

$$-\Delta s \nabla \cdot \mathbf{V}_2 + \langle \bar{\mathbf{V}} \cdot \nabla \bar{s} \rangle + \langle \nabla \cdot \overline{s'V'} \rangle = \bar{F}_{\mathrm{B}}{}^{\mathrm{R}} + \bar{F}_{\mathrm{B}}{}^{\mathrm{S}} - \bar{F}_{\mathrm{T}}{}^{\mathrm{R}} + L\bar{P} \tag{2.8}$$

$$\Delta q \nabla \cdot \mathbf{V}_2 + \langle \bar{\mathbf{V}} \cdot \nabla \bar{q} \rangle + \langle \nabla \cdot \overline{q'V'} \rangle = \bar{E} - \bar{P} \tag{2.9}$$

where P is precipitation. Over land, the flux terms in

(2.7) simplify because $F_B = 0$, assuming zero land heat capacity.

In the tropics, horizontal gradients of temperature and geopotential are small and $\mathbf{V} \cdot \nabla s$ is negligible. This fact is of central importance, as it distinguishes the dynamics of tropical flows from those in midlatitudes. Eddy transports of dry static energy are also small (Oort, 1983; Figs. F25.15–F26.172). A useful approximation to (2.8) is then

$$-\Delta s \nabla \cdot \mathbf{V}_2 \approx L\bar{P} + \bar{F}_B{}^S + (\bar{F}_B{}^R - \bar{F}_T{}^R). \quad (2.10)$$

The time-mean latent heating field has large, sharp maxima while the sensible heat flux is small in most regions. The radiative fluxes provide the relatively small net cooling which tends to cause divergence away from the maxima in P. There is some structure in the radiative cooling associated with convective cloudiness that is correlated with LP but smaller in amplitude. In regions of high time-mean precipitation, the main balance is

$$-\Delta s \nabla \cdot \mathbf{V}_2 \approx L\bar{P}. \quad (2.10')$$

In the moisture equation, one expects $\mathbf{V} \cdot \nabla q$ to be smaller than $q \nabla \cdot \mathbf{V}$ because the mean flow has smaller scales than the mean moisture field. One also hopes that $\nabla \cdot V'q'$ will be small compared with the mean moisture transport although observations are not adequate to make a convincing case. Both of these terms are small over most tropical regions in a GCM moisture budget (N. C. Lau, personal communication, 1986). Therefore,

$$\bar{P} \approx -\Delta q \nabla \cdot \mathbf{V}_2 + \bar{E} \quad (2.11)$$

and in heavily precipitating regions the largest terms are

$$\bar{P} \approx -\Delta q \nabla \cdot \mathbf{V}_2. \quad (2.11')$$

If one averages over a sufficiently large domain, the convergences of dry static energy and moisture will become small, so (2.10′) and (2.11′) can hold only over a small part of the domain of modeling. However, even integrated over an area the size of the GATE A/B array, not all of which is heavily precipitating, moisture convergence still exceeds evaporation by a factor of 2 or 3 (Frank, 1979). Although the balances (2.10′) and (2.11′) represent the (locally) largest terms of (2.10) and (2.11), they cannot form a viable model because they do not relate the mean divergence or precipitation to the forcing or boundary conditions. The apparently smaller flux terms of (2.10) and (2.11) must be retained as well. We believe that the problem is clarified if (2.10) and (2.11) are combined into a moist static energy equation in which the dominant terms are eliminated:

$$-\Delta m \nabla \cdot \mathbf{V}_2 \approx \bar{F}_B - \bar{F}_T. \quad (2.12)$$

Equation (2.12) can also be obtained directly from (2.7) by dropping the terms involving $\mathbf{V} \cdot \nabla m$ and transients.

The quantity Δs plays the role of a static stability in (2.10). In the vertically integrated moist static energy budget (2.12), the same role is played by Δm, which we refer to as the "gross moist stability." We must assume that Δm is a positive definite quantity, so that the time-mean tropospheric adiabatic cooling is always larger than the mean latent heating resulting from low level moisture convergence. Otherwise the time-mean circulation would not be thermodynamically direct. In the case of a zonally symmetric climate, for instance, this constraint is equivalent to the statement that the time-mean Hadley cell must transport energy in the same direction as the flow in its upper branch (to violate this, an atmosphere would have to have extremely strong transport by transient eddies). We emphasize that the time-mean atmosphere can be conditionally unstable, as indeed it is (Riehl and Malkus, 1958) with m decreasing with height at low levels, even though Δm is positive.

Even though Δm must be positive, we expect it to be small in regions of large time-mean precipitation. In fact, while the dominant balances (2.10′) and (2.11′) give us no information about the mean divergence, they do inform us that $\Delta m = \Delta s - L\Delta q \approx 0$ in these regions. Were Δm to become negative, the system would have to adjust, for instance, through an increase in the vertical extent of convection (i.e., an increase in Δs) until the atmosphere were returned to marginal gross stability. Equation (2.12) implies that strong peaks in $\nabla \cdot \mathbf{V}_2$ occur where Δm is small, with low level convergence if the net flux of energy into the atmosphere is positive.

Because the dominant balance has been eliminated, it is more difficult to justify dropping the terms involving $\mathbf{V} \cdot \nabla m$ and $\nabla \cdot m'V'$ to obtain (2.12) than it is to justify (2.10) or (2.11). The most serious problem may be the neglect of transient moisture fluxes as these are known to play a significant role in the local energetics in the subtropics (Oort, 1983; Fig. F26.71, 77, 83, 156, 162, 168). Even though $\nabla \cdot q'V'$ may be small in the moisture budget, we would argue that it plays a significant role in determining the structure of the mean precipitation if it is comparable to $F_B - F_T$ in (2.7).

The balance (2.12) expresses the intuitive idea that horizontal structure in the tropics results partly from horizontal gradients in the flux of energy into the atmosphere through the upper and lower boundaries and partly from horizontal variation in moist stability. Which of these is dominant? In an El Niño year, is the anomalous low level divergence primarily due to anomalous fluxes, $F_B - F_T$, or to anomalous Δm? The fluxes have rather large scales compared to the divergence itself (except for the radiative heating associated with cloudiness, which cannot easily be thought of as the cause of convection, and certain areas of high evaporation in the subtropics, which are not associated with large low level convergence) so structure in Δm must be responsible for most of the sharp horizontal structure in the tropical convergence. A parameterization of Δm provides the basis for the model outlined in the following section.

3. A simple model for the tropical divergence field

We start with the simplified moist static energy budget (2.12) and assume that the energy fluxes through top and bottom boundaries are known or can be parameterized. Any model of F_B, the heat flux into the ocean, will depend on the low level flow itself (through the dependence of evaporation on wind speed in particular), but we assume that the rotational part of the low level flow can be obtained once we know $\nabla \cdot \mathbf{V}_2$, as discussed in the introduction. The problem reduces to modeling Δm.

We begin by assuming that the upper layer moisture, q_1, is negligible, so that $\Delta m \approx \Delta s - q_2$, working in units in which L is unity. Since temperature gradients are small in the tropics, we set Δs equal to a constant. (We shall have to return to this assumption in section 5, however.) We parameterize low level moisture on surface temperature as

$$q_2 = \alpha q_{\text{sat}}(T_s - \delta T) \quad (3.1)$$

where α is a constant less than unity, q_{sat} is the saturation mixing ratio at 1000 mb, T_s is the surface temperature and δT is a constant temperature difference between the surface and low atmospheric levels at which moisture can be thought of as being concentrated. One may think of α as an effective relative humidity. We find that the choices $\alpha = 0.8$ and $\delta T = 1$ K lead to an excellent climatological surface moisture field over the topical oceans (see Appendix). We are assuming in (3.1) that the vertical structure of moisture is sufficiently uniform in the horizontal so that q_2 (or Δq) is proportional to the surface moisture. The end result of these assumptions is the claim that the ratio of the vertically integrated moisture convergence to the low-level mass convergence, Δq, is controlled by the surface temperature, T_s. Combined with the assumption of uniform Δs, this implies that the ratio of the total horizontal energy flux divergence to low level mass convergence, Δm, is controlled by T_s in our parameterization.

The resulting model for low level convergence is

$$-\nabla \cdot \mathbf{V}_2 = \frac{F_B - F_T}{\Delta s - q_2} \quad (3.2)$$

with q_2 given by (3.1). If we define a critical temperature T_c such that $\Delta s = \alpha q_{\text{sat}}(T_c)$, then large low level convergence and precipitation occur when T approaches T_c and the net energy flux into the atmosphere is positive. If one linearizes $q_{\text{sat}}(T)$, then 3.2 implies that $\nabla \cdot \mathbf{V}_2$ is proportional to $(T_s - T_c)^{-1}$ for fixed energy fluxes into the atmosphere.

The sharpness of the convergence zones in this model is determined by how close q_2 comes to Δs. Therefore, the relative magnitude of Δs and the maximum of q_2, denoted $q_{\max}$, is an important parameter. Then $q_{\max}/\Delta s$ should be close to unity; we suspect that any theory for its value would involve the complex interplay between boundary layer and convective heating structure and large scale transients (such as instabilities in the ITCZ). We think of Δs, or T_c, as determined by the constraint that $q_{\max}/\Delta s$ be close to, but not exceed, unity. For an example of how Δs adjusts in a GCM to maintain small Δm as the entire tropics warms or cools (due to changes in the solar constant), see Held and Hoskins (1985, section 2).

The moisture parameterization (3.2) fails in arid regions. One can attempt to define these regions by determining where moisture is required by budgetary constraints to fall below its near-saturation value. Since P cannot be negative, (2.11) suggests the following algorithm:

$$\left.\begin{array}{l} \text{If} \quad E \geqslant \alpha q_{\text{sat}} \nabla \cdot V_2, \quad \text{then} \\ \qquad P = E - q_2 \nabla \cdot V_2 \quad \text{and} \quad q_2 = \alpha q_{\text{sat}} \\ \text{If} \quad E < \alpha q_{\text{sat}} \nabla \cdot V_2, \quad \text{then} \quad P = 0 \\ \qquad \text{and} \quad q_2 = E / \nabla \cdot V_2 \end{array}\right\} \quad (3.3)$$

"Deserts" are defined to be those regions where $P = 0$. In Gill's (1985) recent model, moisture gradients (and, in our terminology, gradients in Δm) only occur in such deserts. We suggest that over most of the tropics, including significant continental regions (such as the Amazon basin and Southeast Asia) and the maritime continent, the criterion for sufficient moisture supply is satisfied for the time-averaged circulation, and that significant gradients in Δm occur because of gradients in q_{sat}. We can avoid the difficulties of applying the desert criterion (which would require a specification of E and would likely have to be modified to include transients and $\mathbf{V} \cdot \nabla q$ terms in the subtropics) by choosing to concentrate on the oceanic convergence zones and simply specifying q_2 over land from observations.

4. Model behavior

Both the numerator and denominator in (3.2) can potentially contribute to the horizontal structure of the model convergence field. Previous simple modeling efforts have focused on parameterizing processes corresponding to the numerator, $F_B - F_T$ (Webster, 1981; Zebiak, 1986; Gill, 1985). Not all of these parameterizations seem to us to be suitable for modeling the climatological convergence fields. We wish to direct attention instead to the effect of the denominator. The simplest means of doing this is to set the numerator, $F_B - F_T$, equal to a positive constant and examine the shape of the convergence zones resulting from the variation in $(\Delta s - q)$ alone. The model convergence field thus depends only on the surface temperature over oceans, and on the prescribed moisture over continents. The results of this calculation, shown in Fig. 1, are only meaningful in the region where the net flux of

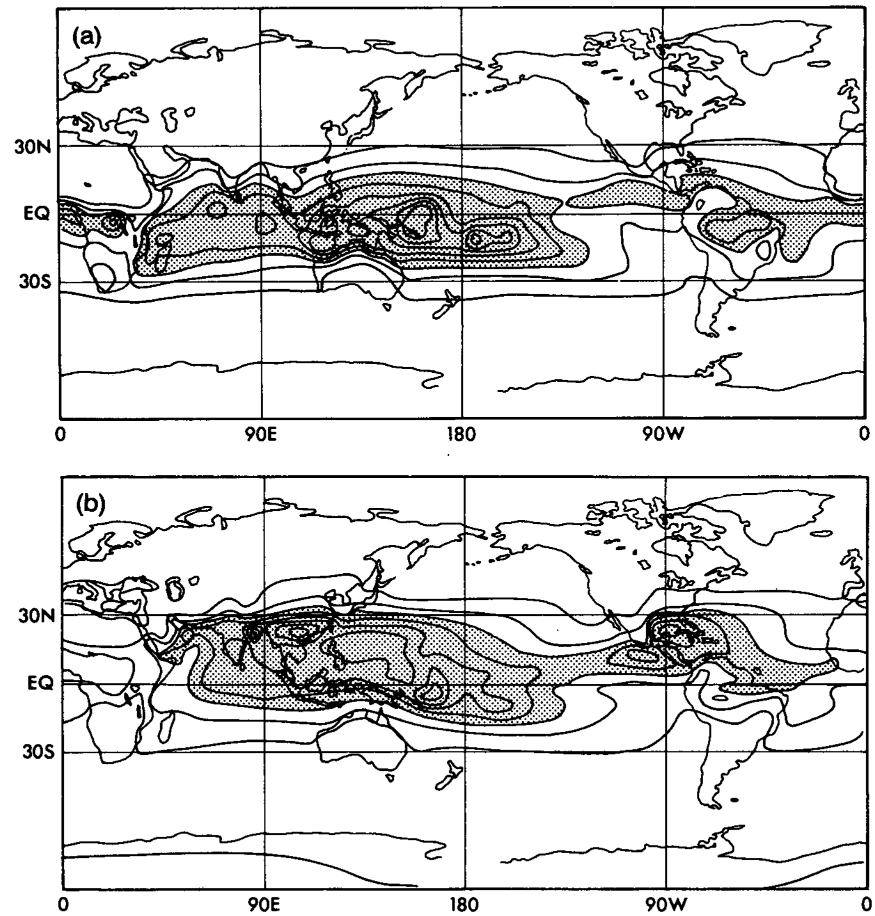

FIG. 1. Model convergence from climatological surface temperature for (a) January and (b) July. Contour intervals may be taken to be 10^{-3} kg m^{-2} s^{-1} (see text for explanation).

energy into the atmosphere is indeed positive, but this is true in the neighborhood of the convergence zones since they fall within the region of largest insolation. In the extratropics, the response has no significance. The parameter $q_{max}/\Delta s$ has been taken to be 0.9 in this case.

Fields of January and July climatological precipitation (Jaeger, 1976) are shown in Fig. 2. Precipitation is chosen as a convenient measure of the convergence zones under the assumption that the balance (2.11′) holds in these zones: regions of large precipitation are also regions of large low level convergence. The location and shape of the convergence zones may be compared with the model results. Maps of model precipitation have almost the same shape as Fig. 1, if we take E in (2.11) to be constant so as to be compatible with our assumption about $F_B - F_T$.

Considering the extreme simplicity of this version of the model, it does well at capturing the gross features of the convergence zones and their seasonal movements. The Indonesian convergence zone, the South Pacific convergence zone (SPCZ), the intertropical convergence zones (ITCZ) in both the Atlantic and Eastern Pacific, and the South/Central American convergence zone are simulated, while the African convergence zone is weak in both seasons and the dry zone separating the SPCZ and ITCZ is not sufficiently well defined in January. The monsoon rainfall over southeast Asia and India in July is present, although the model misses a region of precipitation over the Himalayan Plateau.

Spatial variation in the numerator will modify these results in several important respects. In particular, regions of divergence must be determined by negative regions in the numerator. Unfortunately, it is difficult to specify $F_B - F_T$ from data because the net flux into the ocean, F_B, is poorly known and is likely to be of equal importance to the net flux into the top of the atmosphere, F_T. These effects will be considered in further work, using GCM results as the target climatology to be simulated. Our point here is that the minima in Δm, as parameterized on the surface temperature, appear to contain a great deal of information about the regions of intense convergence.

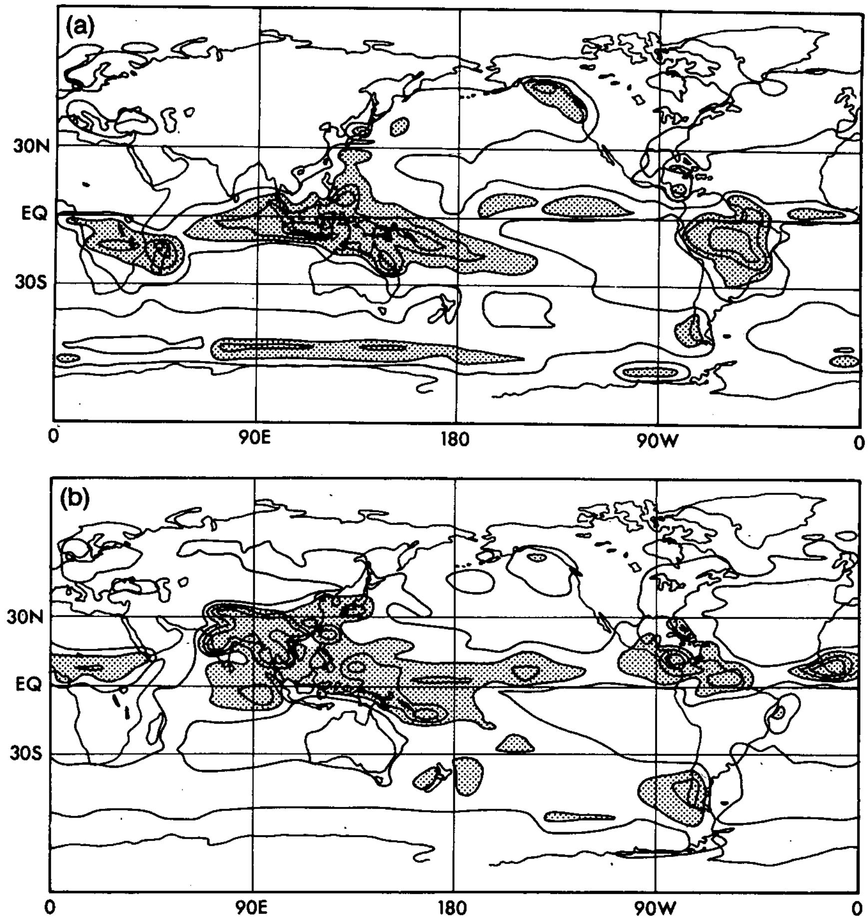

FIG. 2. Climatological precipitation from Jaeger (1976) in mm/month for (a) January and (b) July. Contours at 75 up to 300; stippled over 150.

As a further illustration of this point, Fig. 3 shows the model response to observed sea surface temperatures from the 1972 El Niño event. Over the ocean, monthly mean temperature fields provided by A. H. Oort (personal communication, 1986) are used. Over land, the climatological value of q is used, just as in Fig. 1. The maps for July, 1972, which falls between the "peak" and "transition" phases defined by Rasmussen and Carpenter (1982), and January 1973, which represents the "mature" phase of this event, may be compared to the response to climatological SST in Fig. 1. All parameters are as for the climatological case. In July 1972 the main precipitating zone has moved from Indonesia out to the dateline relative to the climatological response. By January 1973 the region of intense convergence has proceeded even further east to the central Pacific while, over Indonesia, the normal convergence has weakened, since the maximum SST no longer occurs there. In the eastern Pacific, the SST anomalies have comparatively little effect, since the response is a strongly nonlinear function of total SST.

Although we are emphasizing the shapes of the convergence zones, a rough estimate of the magnitudes can be obtained by taking $(F_B - F_T) \approx 50$ W m^{-2} (estimated from a GCM climatology) and $\Delta s \approx 50$ K. This gives a contour interval in Figs. 1 and 3 of about 10^{-3} kg m^{-2} s^{-1}, or 10^{-4} mb s^{-1}. The magnitudes of the convergence zones are within reason, although larger than observed values.

5. Critical comments on the model

While these results seem encouraging, a number of considerations must be borne in mind. If one subtracts the model climatological convergence in Fig. 1 from the ENSO convergence of Fig. 3, the mature phase convergence anomaly is much too weak relative to the divergence anomaly. This suggests that changes in the flux terms may also contribute to ENSO anomalies and it may be desirable to incorporate a parameterization of evaporation, which would require a simple model for the rotational part of the low level flow. The effect of $\mathbf{V} \cdot \nabla q$ could be included in this manner as well. It may also prove necessary to take the effects of eddy moisture flux divergence into account in some

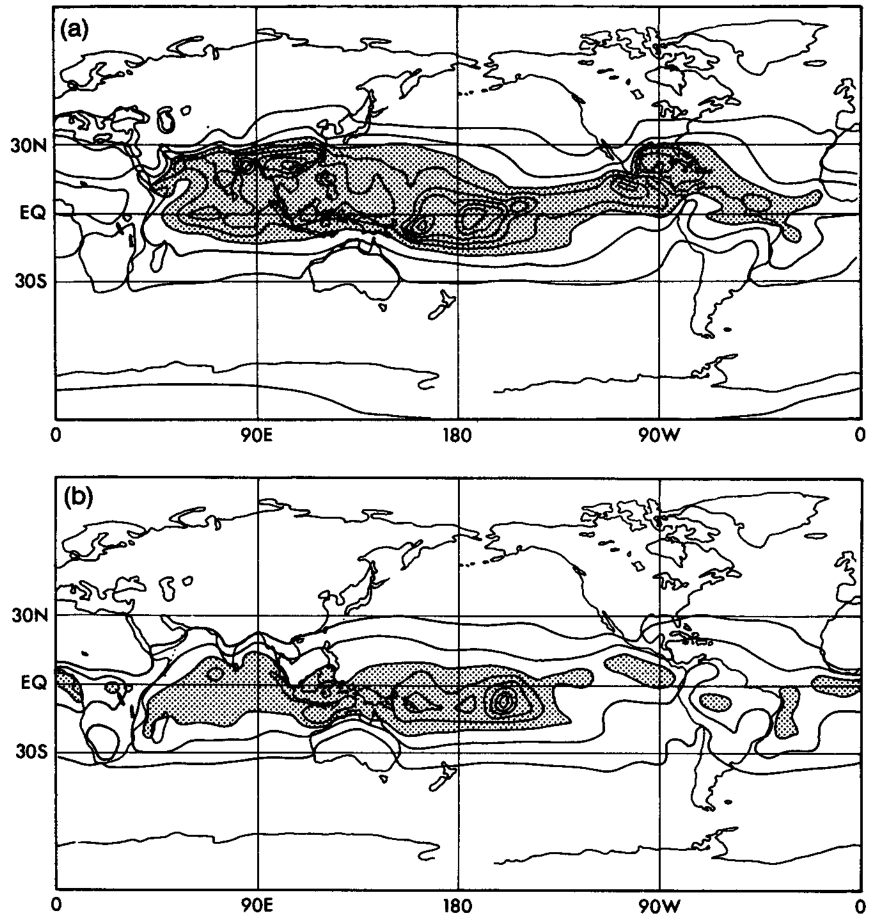

FIG. 3. As in Fig. 1 but for (a) July 1972 and (b) January 1973.

fashion, by specifying them from climatology or by parameterizing them. These would tend to reduce the magnitude of the model convergence zones.

The above appear to us to be relatively straightforward modifications to the model (with the exception of eddy effects). A more serious drawback, since we are making a case for the importance of Δm, is the lack of any theory for horizontal variations in Δs. Use of a constant dry static stability is the usual assumption in a two-layer model. However, our formulation (2.6) makes it clear that variation in the height of convection can affect Δs. If one considers a region of uniformly deep convection with a surrounding region of subsidence then a single value of Δs can be applied everywhere, assuming small temperature gradients, since the upper level outflow from the convective region determines the height of the compensating upper level convergence. However, if there exist variations in the height of convection within a convergence zone, or between different convergence zones, the appropriate specification for Δs is far less clear. The weakness of the African and South American convergence zones relative to the Indonesian convergence in Fig. 1 may well be due to such variations. As described in section 3, we would like to think of Δs as determined by the maximum low level moisture. But to what extent does the convection over Indonesia help determine the effective Δs over the Amazon?

Finally, our formulation of a two-layer model is motivated by and depends on our assumption of a simple vertical structure for the divergence. If this structure is more complex (e.g., over deserts where sensible heating is balanced by upward motion at low levels, despite upper level subsidence) the divergence weighted integrals in (2.6) will not be physically useful and may, in fact, be undefined, since the zeroes in $\nabla \cdot \mathbf{V}_2$ and $\langle m \nabla \cdot \mathbf{V} \rangle$ need not coincide. Thus, while it may be useful to think of a background Δs and Δm in a two-layer model, there are limits to how far the analogy can be carried in a continuous atmosphere or multilevel model. In particular, wherever $\nabla \cdot \mathbf{V}_2$ is small, the definitions (2.6b,c) can lead to a noisy Δm field; the value of our two-level model depends on these regions being dynamically unimportant.

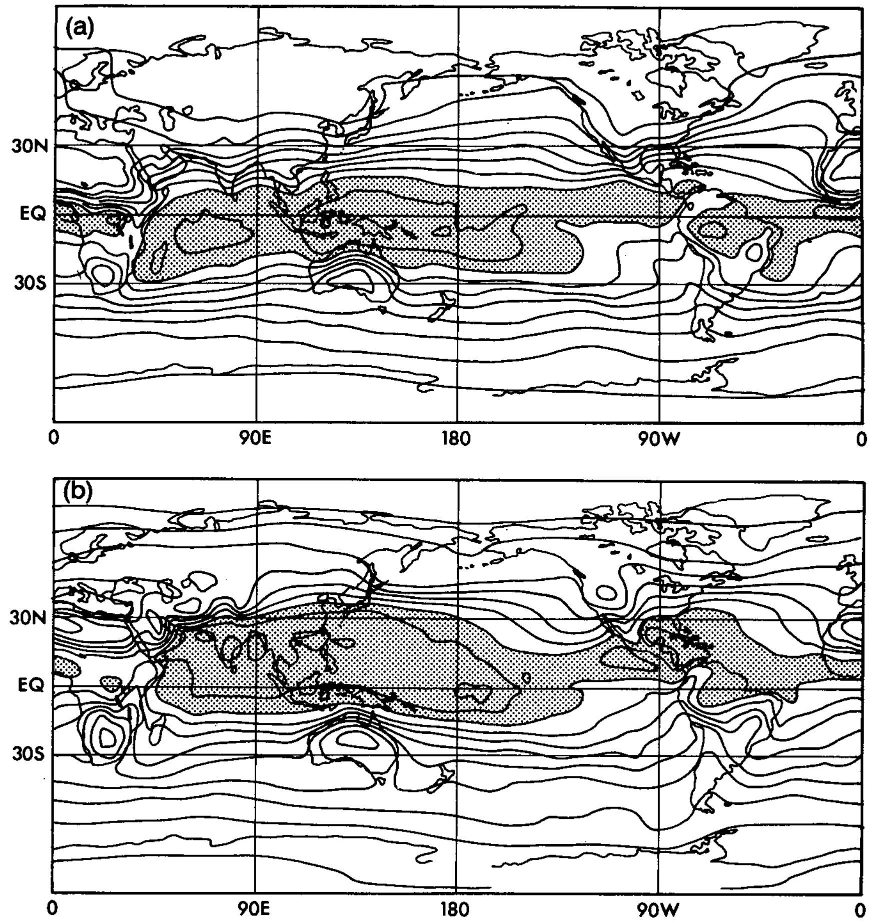

FIG. 4. Climatological surface moisture from Oort (1983) in temperature units (conversion factor L/C_p): (a) January; (b) July. Contours at 5 deg; stippled over 40 deg.

6. Conclusions

The problem of relating tropical vertical motion and precipitation to the lower boundary conditions has been examined using a model based on the vertically integrated moist static energy equation. A two-layer formulation arises naturally from assumptions about the vertical structure of the divergence, and suggests that a vertically integrated measure of the moist static stability, which we have called the gross moist stability, is of crucial importance in determining the form of the convergence zones. An extremely simple version of the model was presented which proved capable of reproducing *qualitatively* the positions and movements of the tropical convergence zones, both for the climatology and for ENSO events. The model makes more precise the idea of a critical surface temperature required for sustained deep convective activity. Warmer surface temperatures create mean upward motion by increasing the low level moisture and thereby increasing the instability of the flow to moist convection. Precipitation in the favored regions is enhanced by convergent moisture transport at the expense of the less favored regions.

The model may help resolve an old controversy typified by Ichiye and Petersen (1963) and Bjerknes (1969) on the one hand, and Ramage (1977) on the other. The former note the correlation between rainfall and SST and assume it is due to enhanced latent and sensible heat fluxes in regions of warm SST. Ramage takes issue with this, since these fluxes do not correlate with SST or rainfall in the manner required by the hypothesis. In our model, high SST does indeed cause high rainfall, but no change in the heat flux into the atmosphere is required (although variations in the heat flux may occur, combining their effects with that of the moist static stability). The case presented here illustrates this by having variation in model precipitation due entirely to moisture convergence. This convergence depends nonlinearly on SST through the effect of SST on the gross moist stability.

While this "minimal" model suggests which processes are important to include in current simple modeling efforts, it also raises questions about the limitations of such models. In particular, the sharp structure of the convergence zones must come from horizontal variation in the moist static stability. However, the pa-

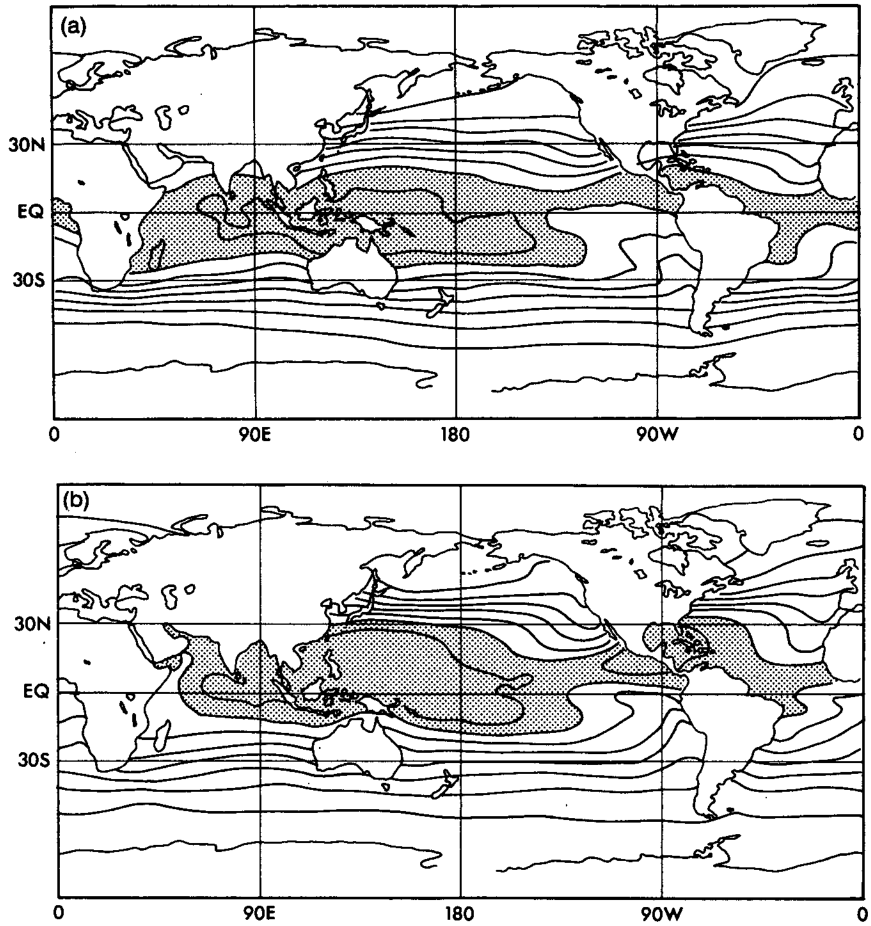

FIG. 5. Moisture parameterized on climatological sea surface temperature from Oort (1983): (a) January; (b) July. Units and contouring as for Fig. 4.

rameterization of this quantity involves broad assumptions since the effects of many complex convective processes must be condensed into a few parameters. The results of our parameterization seem encouraging but it is not clear whether they can be made more quantitative.

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APPENDIX

Parameterization of Moisture on Surface Temperature

Over most of the tropics there is a plentiful supply of moisture and on average the air at low levels tends to be as near saturation as boundary turbulence allows. Furthermore, the temperature at low levels is closely tied to the surface temperature (Oort, 1983; Figs. F103–153). These considerations suggest the parameterization (3.1), which we test here against the climatological surface moisture field. The "World Weather Records extended" surface mixing ratio analysis from Oort (1983, pp. 5–8) is shown in Fig. 4. Using sea surface temperature from the Oort dataset, and setting $\alpha = 0.8$ and $\delta T = 1$ K, the parameterized moisture field in Fig. 5 is in excellent agreement with the observed field over tropical oceans. A similar comparison, using a near-surface temperature analysis which has good data coverage over both land and ocean, suggests that (3.1) remains valid over significant regions of the tropical continents, although it obviously fails in arid regions, such as the Sahara, the Middle East and over the southwestern United States in July and Australia in January.

The moisture in Figs. 4 and 5 is expressed as its latent energy equivalent in units of temperature (conversion factor L/C_p). The horizontal gradients in this field (e.g., the 15 deg difference across the Pacific) are much larger than those which occur in the temperature field.

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