Simulation of Tropical Climate with a Linear Primitive Equation Model*

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ABSTRACT

The tropical climate simulated with a new global atmosphere model is presented. The model is purposely designed for climate studies and is still under development. It is designed to bridge the gap between very efficient but simple models of the tropical atmosphere and sophisticated but inefficient general circulation models (GCMs). In this paper the authors examine the sensitivity of the model's climate to specific formulations of convection, boundary-layer physics, and radiation.

The model uses the Betts–Miller convection scheme and a parameterization of the planetary boundary layer (PBL) that combines similarity theory for computation of surface fluxes with a simple scheme for diagnosing PBL depth. Radiative cooling is specified and land surface processes are bypassed by relaxing modeled low-level values to observed quantities. Orography is ignored. The model contains six vertical layers and has a horizontal resolution of about 3° × 5.625°.

The authors compare the climate simulated with two different versions of the Betts–Miller convection scheme. More realistic simulations of rainfall are obtained with the later version, which includes the effects of convective downdrafts. These, by cooling and drying the PBL, act to restrict the areas of convection while strengthening the intertropical convergence zone. The sensitivity to choice of PBL physics is less, and quite similar results were obtained when the PBL scheme was replaced with constant exchange coefficients and PBL depth. In contrast, the amount of precipitation varied strongly with the prescribed radiative cooling. The important role that shallow convection and cloud-radiation interactions play in the spatial organization of deep convection is demonstrated, by default, in an experiment using clear-sky radiative transfer.

The modeled climate, as judged qualitatively by its simulation of quantities of importance to air–sea interaction and climate, such as the low-level wind field and precipitation, is in many ways comparable to that achieved by much more complex GCMs. Indeed the rainfall simulation appears better than obtained by many models that use other convection parameterizations. This adds to the accumulating evidence that the Betts–Miller scheme is a quite reliable scheme, at least for simulation of convection in the current climate. A major model flaw is a very poor Asian summer monsoon, which is attributed to lack of orography in the model. It is demonstrated, by inclusion of a specified monsoonal forcing, that this also has an effect, though modest, on the simulation of the trade winds over the Pacific.

The results suggest there is hope for development of models of intermediate complexity that achieve a degree of realism exceeding the simple models that have often been used in El Niño studies while retaining much of their efficiency.

1. Introduction

Climate research has been transformed over the last several years by the demonstration that prediction of certain aspects of climate variability, with lead times of up to a year, is now possible. To date, predictive skill has been demonstrated only in regard to the El Niño/Southern Oscillation (ENSO) (Cane et al. 1986; Latif et al. 1994). Routine dynamical forecasts of ENSO are now made using a coupled atmosphere–ocean model of the tropical Pacific (Cane 1991). Forecasts from this model are taken into consideration during the economic and agricultural decision making process by several affected countries, for example, Peru and Brazil (Moura et al. 1992). While dynamical predictions have so far been limited to the tropical Pacific sector, statistical correlations exist between ENSO and worldwide weather fluctuations (Ropelewski and Halpert 1987, 1989). Teleconnections exist that transmit the influence of anomalous conditions in the tropical Pacific into midlatitudes where they can contribute to seasonal variability (Wallace and Gutzler 1981). On longer timescales the observed decadal scale variability of ENSO has been hypothesized to be related to similarly long timescale variability in midlatitudes (Graham 1994).

The predictability of ENSO raises the question of whether these related phenomena are also predictable in some degree. In addition the question must be ad-

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dressed as to whether other tropical climate phenomena, such as African and Brazilian drought and Asian monsoon variability, each of which may be linked to variability of sea surface temperature (e.g., Palmer et al. 1992), are also predictable. It is easy to extend the list to other phenomena such as the observed decadal scale variations in North Atlantic climate (Kushnir 1994).

Assessment of predictive skill requires models. The length of the model integrations that are required is already quite long, and, as we begin to consider longer timescale changes, it will become still longer. These requirements make operational climate prediction with conventional coupled atmosphere–ocean GCMs prohibitively expensive. Attempts to shortcut these requirements are in danger of error. For example the knowledge and experience that underlies the success (such that it is) of the prediction effort at Lamont-Doherty Earth Observatory rests on coupled model integrations that have now extended, in total, over many tens of thousands of simulated years. It is only through the analysis of such extended integrations that the nature of the model oscillation and its sensitivity to uncertain parameters could be understood (Zebiak and Cane 1989).

This raises an immediate problem. The coupled model includes only the Pacific sector. It cannot predict the hypothesized connections with midlatitude climate or even to other tropical regions. It also cannot include the effects of other regions (e.g., the Indian Ocean sector and the Asian monsoon) on the Pacific. Dealing with these issues requires a global model but existing global atmosphere models are currently too computationally slow to conduct the very long experiments that are required to address reliably issues of global climate variability and predictability. This will be less of a problem as computer technology advances but will remain a real constraint for the forseeable future.

With this in mind we have sought to develop a global atmosphere model that overcomes many of the problems encountered with simpler models such as those of Zebiak (1986) or Seager (1991) while retaining an efficiency that allows long-term integrations. Previous simple models were based on the familiar Gill (1980) model for a steady-state, thermally forced circulation on an equatorial beta plane, which is subject to Rayleigh friction and contained within a single baroclinic vertical mode. The Gill model has proven capable of simulating the low-level wind anomalies associated with ENSO (e.g., Zebiak 1986). It is, however, far less appropriate to the other problems listed above. Its limitations can be listed as follows [see Seager (1991) for a more detailed criticism]. First, its equatorial nature means it cannot be applied to problems that involve tropical–midlatitude interactions. Second, the single vertical mode, coupled with the need to simulate near-surface winds, necessitates that the entire atmosphere be subjected to a large damping that prevents propagation of signals over large distances, making studies of even tropical teleconnections impossible. Third, the simple vertical structure means that both radiation and convection must be treated very simplistically, and, while this has apparently mattered little in the ENSO prediction work, it is certain to become a problem as we attempt to improve forecasts and look at other phenomena. The simple vertical structure also seriously restricts the treatment of the atmospheric boundary layer and surface fluxes of heat and momentum. Motivated by these concerns, we have constructed a new global atmosphere model that aims to fall between the simple models and current GCMs but that, if desired, is generalizable to the level of complexity of a GCM.

The dynamical framework of this model was introduced in Seager and Zebiak (1994, SZ94 hereafter) where, in idealized form, it was applied to a theoretical problem concerning interactions between convection and dynamics. The novelty of the model is expansion of the primitive equations on a sphere into model normal modes. This reduces the time integration to a set of ordinary differential equations governing the amplitude of each normal mode in terms of its frequency and its forcing. If the forcings (physics, nonlinear advections, etc.) are known, the ODEs can be advanced analytically without introduction of computational dispersion or phase errors, and the integration will be stable for any length of time step. This allows long time steps; we currently use 1 day. Although the actual computation per time step is slightly less than for a traditional spectral model, the major gain in efficiency is through the ability to choose the time step based on accuracy considerations rather than by the need to avoid numerical instability. Model equations and an extensive discussion of the model numerics and architecture can be found in the appendix of SZ94.

Our ultimate goal is to simulate a realistic climatology and climate variability. We present here our first attempt to do this in a simulation of the tropical seasonal cycle. At the present stage of development the goals are still limited. For example, we do not attempt to simulate any midlatitude atmospheric phenomena at all. This is partly because the more immediate interests are in the Tropics but also because the midlatitude climatology is quite nonlinear and, at present, we have included only a handful of nonlinear terms. Exclusion of nonlinear horizontal advections also means we cannot simulate a realistic Hadley cell in the Tropics, so we expect our zonal mean flow to be poor. Further, even in the Tropics, teleconnections occur through interactions between the basic state and anomalous circulations. In the absence of the correct zonal mean basic state the model will not be able to reproduce observed tropical teleconnections (Webster and Chang 1988). We expect this to be more of an issue in reference to interannual variability and restrict ourselves here to the tropical climatology.

We also have chosen to enhance the model's realism in a gradual way. This is of course practical but also has the advantage of allowing examination of the contribution of different physical processes to the model climate. At this stage we include quite sophisticated treatments of both convection and the planetary boundary layer. On the other hand, the treatment of radiation and land surfaces are highly idealized and there is no topography. The latter exclusion immediately brings into question whether the model will be able to produce a summer Asian monsoon [the quality of simulation of which has been linked to topography going back to Hahn and Manabe (1975)], and we will see that the model does indeed do a poor job of this. However, in regions of less topographic influence, we will see that this "minimal physics," linear, tropical model can reproduce with a certain skill many features of the observed climatology, including aspects of the distribution of precipitation. To some extent the errors we encounter are of the same magnitude as those found in less efficient but more sophisticated models, which raises the hope that ultimately this model will be a highly useful and practical model with which to study climate variability.

This paper is best read as a report on "a model in progress," and we provide an extensive description of the model physics. Our own understanding of the tropical atmosphere has been enhanced by our efforts to improve the model's realism. Although most of the results we present are acceptable by common standards, throughout we will talk about what we found *did not* work. This is because we learned as much from our errors as we did from our successes, and we believe that such an approach will best serve those in the modeling community actively involved in building models of intermediate complexity. We will present and discuss results from a standard simulation of the model in its most complete form and compare them with simulations that contain different or simpler parameterizations of either radiation, convection or PBL. We believe this is an effective way to assess the importance of various processes in establishing the observed tropical climate.

2. Model description

a. Dynamics

Since SZ94 provide an extensive description of the model dynamics, only a brief review will be provided here. The primitive equations on a sphere are linearized around a resting basic state with temperature a function of the vertical coordinate only. The vertical coordinate is $\sigma = p/p_s$, where p is pressure and p_s is surface pressure. The equations are vertically discretized using an energy conserving form, and an algebraic vertical structure equation is formed. The vertical structure equation is solved numerically to yield as many vertical modes and associated equivalent depths as there are layers in the discretized system. The vertical differencing is chosen such that the vertical modes are orthogonal.

Expanding the primitive equations in terms of these vertical modes transforms the system into a set of two-dimensional shallow water equations. The shallow water equations are then expanded in terms of spherical harmonics. Truncating the expansion again yields a matrix eigenvalue problem. In this case the eigenvectors can be used to construct two-dimensional patterns of velocity and pressure that correspond to a normal mode, or Hough mode, of the shallow water equations on a sphere. The corresponding eigenvalue is the frequency of the mode. The modes divide into gravity waves, Kelvin waves, Rossby waves, and mixed Rossby–gravity waves. The modes are orthogonal whatever the truncation.

The structures of these normal modes are discretized onto a Gaussian grid of a resolution chosen to preserve the orthogonality of each mode to all the others and saved. They form a complete basis set for construction of solutions. This three-dimensional spectral expansion reduces the entire system to a set of uncoupled ordinary differential equations governing the amplitudes of the normal modes. The equations can be solved analytically without introduction of phase errors or computational instability, whatever the time step used, providing the forcing on the modes is known and is fixed for the duration of the time step. This provides the main computational advantage of our method. It allows the time step to be chosen on accuracy and application considerations rather than being dictated by computational stability considerations.

The algorithm for construction of solutions is then as follows. Beginning in physical space, three-dimensional forcing fields (radiation, convection, friction, nonlinear terms, etc.) are computed. These are projected onto the three-dimensional modes. This involves simple dot products for the vertical and meridional transforms and Fast Fourier Transforms in the zonal direction. The projection provides the forcings on each mode that are used to update the modal amplitude equations. Physical space fields are then derived by multiplying each modal amplitude by its normal-mode structure and summing over all the modes. Then the forcing fields are recomputed in physical space beginning the next model cycle.

It is important to remember that while we use the normal modes of the linear primitive equations as a basis set, this method is capable of representing fully nonlinear solutions. Indeed, during initialization of weather-prediction models, the fully nonlinear state of the true atmosphere is approximated through use of linear model normal modes. Hence our method is applicable to both linear and nonlinear circulations. There remains the issue, however, of what particular set of normal modes will perform best in any particular situation (see Errico 1991).

The model has six vertical layers. The upper four have sigma thickness of $\delta\sigma = 0.2$, and the lower two have thickness $\delta\sigma = 0.1$. The horizontal resolution corresponds to about 3° × 5.625°.

b. Convection

We have adopted the scheme introduced by Betts (1986) and Betts and Miller (1986, hereafter BM86). The motivation for the scheme is the observation that there is a quasiequilibrium between the tendencies of convection to stabilize the atmosphere and of large-scale forcing (radiation, surface fluxes, and motions) to destabilize it (Betts 1986; Arakawa and Schubert 1974). If the model vertical column becomes unstable, then the model temperature and humidity profiles are relaxed back to empirically derived quasi-equilibrium profiles over a small, but finite, timescale. Hence the convective forcings of temperature, T, and humidity, q, are given by

$$\left(\frac{\partial T}{\partial t}\right)_{\text{conv}} = \frac{T_{\text{ref}} - T}{\tau}, \tag{1}$$

$$\left(\frac{\partial q}{\partial t}\right)_{\text{conv}} = \frac{q_{\text{ref}} - q}{\tau}, \tag{2}$$

where the subscript ref. refers to the reference profiles and τ is the relaxation timescale, taken to be a few hours. The values of T_{ref} are obtained by inverting reference profiles of saturation equivalent potential temperature, θ_{ES}, and q_{ref} is then obtained from T_{ref} assuming a profile for the pressure difference between the model-level pressure and the saturation-level pressure for air with temperature T_{ref}. This pressure difference is referred to as the subsaturation.

Stability is assessed relative to a moist virtual adiabat defined as one that retains its condensed water. The use of a virtual adiabat follows from the observation of Betts (1982), later confirmed by Xu and Emanuel (1989), that the atmosphere is close to neutrally stable to such adiabats while being very unstable to a moist pseudoadiabat in which condensed water is rained out. The argument follows that it is the former that should be more appropriate to the lower atmosphere. If this is so, then the conditional instability of the lower atmosphere is less than once thought, which removes the need to search for suppression mechanisms and "triggers" for convection.

For deep convection the reference profile is marginally unstable to a moist virtual adiabat from cloud base to the freezing level. Above freezing level the reference profile is found through linear interpolation in pressure between θ_{ES} at the freezing level and the value at neutral buoyancy (cloud top). This can be regarded as an approximation to the ice adiabat. For shallow convection the procedure is quite different. The reference profile values for T and q are assumed to lie along a mixing line between mixed-layer air at cloud base and the air above the inversion (Betts 1982).

An important feature of the BM86 scheme is that it conserves moist static enthalpy. For deep convection, after the first-guess reference profiles of T_{ref} and q_{ref} have been established, the total enthalpy constraint

$$\int_{\sigma_b}^{\sigma_t} (k_{\text{ref}} - \bar{k})\, d\sigma = 0 \tag{3}$$

is imposed, where $k = c_pT + Lq$, σ_b and σ_t are the σ values at cloud base and cloud top, and the overbar denotes the unadjusted thermodynamic state. To achieve this, the quantity

$$\Delta k = \frac{1}{\sigma_b - \sigma_t} \int_{\sigma_b}^{\sigma_t} (k_{\text{ref}} - \bar{k})\, d\sigma \tag{4}$$

is calculated, and T_{ref} is adjusted at each level by an amount that changes k_{ref} by Δk while maintaining the subsaturation at the same prescribed amount. Since the first-guess reference profile is constructed through the low-level equivalent potential temperature, θ_E, it leaves the lowest-level temperature unchanged. It is the requirement for conservation of moist static enthalpy that adjusts the reference profiles and forces drying and cooling of the boundary layer.

Deep convection does dry and cool the boundary layer (e.g., Emanuel 1994). However, the way that the BM86 scheme achieves this is somewhat arbitrary. In reality it occurs by convective downdrafts that introduce into the boundary layer cold and dry air from above. Recently Betts and Miller (1993, hereafter BM93) have modified their original scheme to account for reduction of boundary-layer enthalpy by unsaturated downdrafts. The modification involves two changes to the original scheme. First, the first-guess reference profile of layers below 850 mb are calculated as those that correspond to a downdraft originating at 850 mb and with the properties of environmental air at that height and moving down at constant θ_E and with constant subsaturation. The reference temperature and humidity in the boundary layer are thus coupled to drier and colder air above in accord with expectations. Second, a longer adjustment time, τ_{PBL}, is used for the boundary-layer convective adjustment. BM93 coupled this timescale to the precipitation and evaporation into the downdraft as

$$\frac{1}{\tau_{\text{PBL}}} = \frac{\alpha \rho_w g P}{p_s \int_{\sigma_b}^{\sigma_{\text{BL}}} \Delta q_c\, d\sigma}, \tag{5}$$

where Δq_c is the difference in humidity in the downdraft relative to its inflow value, ρ_w is the density of water, σ_{BL} is at the downdraft inflow level, and the precipitation, P, is in units of meters per second. Here

α is a constant taken by BM93 to be 0.25. The adjustment time hence reduces with the intensity of convection. We found this led to unrealistic fluxes in convecting regions and instead adopted a fixed value of $\tau_{\mathrm{PBL}} = 0.5$ days. To account for the differing timescales of adjustment, the energy conservation constraint needs to be modified to

$$\Delta k = \frac{\tau}{\sigma_b - \sigma_t} \left\{ \int_{\sigma_b}^{\sigma_{\mathrm{BL}}} \left[\frac{(k_{\mathrm{ref}} - \bar{k})}{\tau_{\mathrm{PBL}}} \right] d\sigma + \int_{\sigma_{\mathrm{BL}}}^{\sigma_t} \left[\frac{(k_{\mathrm{ref}} - \bar{k})}{\tau} \right] d\sigma \right\}. \quad (6)$$

The precipitation is given by

$$P = \frac{p_s}{g \rho_w} \left\{ \int_{\sigma_b}^{\sigma_{\mathrm{BL}}} \left[\frac{(\bar{q} - q_{\mathrm{ref}})}{\tau_{\mathrm{PBL}}} \right] d\sigma + \int_{\sigma_{\mathrm{BL}}}^{\sigma_t} \left[\frac{(\bar{q} - q_{\mathrm{ref}}}{\tau} \right] d\sigma \right\}. \quad (7)$$

Shallow convection is assumed to be nonprecipitating. Consequently, two separate energy constraints are imposed:

$$\int_{\sigma_b}^{\sigma_t} (T_{\mathrm{ref}} - \bar{T}) d\sigma = \int_{\sigma_b}^{\sigma_t} (q_{\mathrm{ref}} - \bar{q}) d\sigma = 0. \quad (8)$$

This is achieved by computing quantities analagous to that in (4) but separately for T and q, and then adjusting the reference profiles by these amounts at each level between cloud base and cloud top. Note that in contrast to deep convection, the shallow convection scheme determines its own subsaturation values.

The Betts scheme is attractive because of its conceptual simplicity and reasonable numerical efficiency (it was coded in house and uses several look up tables to speed computation of the thermodynamic profiles) and the fact that within it convection occurs solely in response to bouyancy and not extraneous factors such as moisture convergence as in the once popular scheme of Kuo (1974). It is open to criticism because the universality of the reference profiles has not been established [see Bretherton (1993) for a look at the shallow convection case].

Both schemes have been used in atmosphere models. Slingo et al. (1994) and BM93 used the BM93 version in the European Centre for Medium-Range Weather Forecasts (ECMWF) model. Janjić (1990, 1994) has used the BM86 version in the National Meteorological Center's eta model and credited the scheme with improving precipitation forecasts. The precipitation simulations presented for the ECMWF model are, at least in our opinion, much improved over earlier versions that used the Kuo scheme. The BM86 scheme has also been used in tropical cyclone simulations by Baik et al. (1990). Here we will present simulations with both versions of the convective parameterization.

The convective relaxation times are shorter than the model time step. We perform a one-dimensional radiative–convective calculation, with a time step of a few hours, at each model grid point in order to derive convective forcings averaged over the dynamical model time step. More details can be found in SZ94.

c. *Planetary boundary layer*

Most current GCMs attempt to simulate processes in the planetary boundary layer (PBL) through placement of a large number of model layers close to the ground, a similarity theory to calculate the surface fluxes, and some parameterization of the turbulence in the PBL (see Sommeria 1988 for a review). It is hoped that this approach will yield a mixed, or partially mixed, PBL as observed. A different approach is that of Suarez et al. (1983), which a priori assumes the existence of a mixed layer and calculates its depth using a turbulent kinetic energy closure. In the former approach the lowest model level is typically very close to the surface so only a similarity theory for the surface layer is invoked. This allows calculation of the exchange coefficients that appear in the formulas for the surface fluxes of heat and momentum in terms of quantities at the lowest model level (e.g., Louis 1979).

Our model has considerably coarser vertical resolution than most GCMs, so that our lowest model level is around 950 mb—well within the turbulent part of the PBL that lies above the surface layer. In the early days of atmospheric GCMs coarse vertical resolution was also prevalent, and this problem was addressed by matching a similarity theory for the surface layer with one for the turbulent layer. The most commonly used such scheme is that of Deardorff (1972), and we adopt it here.

Deardorff's scheme contains a number of formulas for calculating the exchange coefficients in terms of the PBL depth and a Richardson number evaluated over the PBL depth. The formulas are analytical approximations to the experimental data of Businger et al. (1971). It uses a free-convection limit to calculate fluxes at low wind speeds. The scheme also allows calculation of the direction of the surface wind stress. It is assumed that right at the surface the friction is sufficiently large that the pressure gradient balances the stress. Once the magnitude of the stress is known, then a simple geometric exercise is performed to find the direction of the pressure gradient that will balance this stress. This is then the direction of the surface wind stress.

Deardorff (1972) suggested a prognostic equation for the calculation of the PBL depth. We instead opt for the simpler scheme of Troehn and Mahrt (1986), which calculates the PBL depth, h, in terms of a critical bulk Richardson number, Ri_c, from

$$h = \mathrm{Ri}_c \frac{T_0 |v(h)|^2}{g[\theta_v(h) - \theta_s]}. \quad (9)$$

Here T_0 is the PBL mean temperature, $v(h)$ is the velocity at h, $\theta_v(h)$ is the virtual potential temperature at h, and θ_s is the surface virtual potential temperature. The model level for which $\mathrm{Ri} \geq \mathrm{Ri}_c$ is found, and h is then found by linear interpolation between this Richardson number and the one below. This scheme provides qualitatively reasonable PBL depths in the experiments we ran, with the PBL top varying from about 930 mb over the colder parts of the tropical oceans to 700 mb over the Sahara during northern summer. The scheme has also been implemented in the National Center for Atmospheric Research Community Climate Model by Holtslag and Boville (1993).

Troehn and Mahrt (1986) and Holtslag and Boville (1993) both use similarity considerations to simulate the mixing in the PBL and include countergradient heat fluxes. Again, we use a simpler approach. We calculate the mass-weighted mean potential temperature and humidity over the computed PBL depth and relax the model values in the included levels to the means over a timescale of a few days. Since the PBL depth is typically within a model layer, care must be taken to properly weight these tendencies. This mixing moves moisture upward but, since the potential temperature typically increases with height, moves heat down and assists in creating an inversion at the PBL top.

The simple case with which this scheme will be compared uses fixed exchange coefficients and PBL depth and calculates stress with the lowest model level winds, much like SZ94. In that case the surface stress and lowest model level wind are necessarily in the same direction.

d. Radiation

Many simple atmosphere models parameterize radiative cooling by relaxing the model temperature to an assumed equilibrium profile. For example, Schneider and Lindzen (1977) relax temperature to a dry adiabatic lapse rate constructed through the surface temperature, a formulation that has been commonly used since. This provides a vertical profile of radiative cooling with maximum cooling in the upper troposphere and near-zero cooling near the surface. In contrast, radiative transfer calculations produce cooling rates that maximize near the surface and decrease to the tropopause (e.g., Ramanthan and Downey 1986). Such a profile is easily understood in terms of the concentration of moisture near the earth's surface. Seager and Zebiak (1994), recognizing that the near surface radiative cooling is a crucial component of the atmospheric thermodynamics that govern the location of convection, modeled the radiative cooling as

$$Q_{\mathrm{rad}} = R - T'/\tau_{\mathrm{RAD}}, \tag{10}$$

where R is an imposed radiative cooling profile that peaks near the surface and is zero at the top of the atmosphere. The other component damps the model temperature anomaly T' on a timescale, τ_{RAD}, and is considerably smaller. Term R was taken to be zonally uniform and to decay in amplitude away from the equator. Although this was qualitatively correct, it is extremely crude and ignores any effects of cloud. However, full radiative transfer calculations are expensive and would add considerable complexity to the model. Hence, in this work, we retain this parameterization but extend it to account, in a similarly crude way, for the effects of clouds associated with deep and shallow convection. We do this by making the radiative cooling profile R depend on cloud cover and the convective regime

$$R = (1 - f_c)R^{\mathrm{clr}} + f_c R^{\mathrm{cld}}, \tag{11}$$

where f_c is the fractional cloudiness and is taken to be 0.6 if convection is present and zero otherwise, R^{clr} is the clear-sky radiative cooling profile and R^{cld} is the cloudy-sky profile. At the surface the clear-sky cooling is 2.66 K day^{-1}, increasing to 4 K day^{-1} at $\sigma = 0.85$, and then decreasing to zero in the top model level. It is also assumed that above $\sigma = 0.8$ the clear-sky cooling is one-half of these values if no deep convection is present. This is supposed to account for the much smaller values of moisture in the subsiding areas of the Tropics.

Machado and Rossow (1993) have calculated the total (solar plus longwave) cooling profiles in deep convective clouds. Their results suggest that below cloud top the radiative cooling can be crudely assumed to approximate zero but that strong radiative cooling occurs at cloud top. A similar process occurs in regions of shallow convection. In the latter case the cloud-top cooling has long been recognized to be important to destabilization of the shallow convective layer and maintenance of the capping inversion. Cloud-top radiative cooling plus evaporative cooling balance the warming due to subsidence in the inversion (Betts and Ridgway 1989). This recognition goes back at least as far as Lilly (1968), and Lilly and Schubert (1980) discuss some early models of radiatively driven convective boundary layers. This instability is also included in the model of the trade wind boundary layer introduced by Albrecht et al. (1979) and Albrecht (1979) and later interpreted by Bretherton (1993). Although differences exist between the models, they all assume that within the cloudy portion of the sky the radiative cooling is contained within a thin layer near the cloud top and is zero below. It is also commonly assumed that the *vertically integrated* radiative cooling is the same in the cloudy and clear portions of the sky. These are the assumptions we adopt for both deep and shallow convection.

First we calculate the radiative cooling, $\bar{R}$, integrated (in discretized form) from the cloud top, σ_{ct}, to the surface, which is given by the clear-sky profile as

$$\bar{R} = [(\sigma_{l_{ct+1/2}} - \sigma_{ct})R_{l_{ct}}^{clr} + \sum_{l=l_{ct}+1}^{L} R_l^{clr}\delta\sigma_l], \quad (12)$$

where σ_l is the sigma value at the middle of level l, $\sigma_{l+1/2}$ is the value at the base of level l, $\delta\sigma_l$ is the thickness of level l, l_{ct} is the level containing the cloud top, and L is the total number of model levels. Then the radiative cooling profile for convecting columns is

$$R^{cld} = \begin{cases} R^{clr}, & \text{if} \quad l \leq l_{ct} \\ [R^{clr}(\sigma_{ct} - \sigma_{l_{ct}-1/2}) + \bar{R}]/\delta\sigma_{l_{ct}}, & \text{if} \quad l = l_{ct} \\ 0, & \text{if} \quad l \geq l_{ct}. \end{cases} \quad (13)$$

This radiation parameterization attempts to crudely account for some aspects of cloud radiation interaction, although it should be noted that the magnitude of the cooling is set somewhat arbitrarily. The general pattern, allowing for cloud cover, is in rough agreement with those shown by Newell et al. (1974) and Morcrette (1990) but can only be justified as an interim procedure before adoption of a radiative transfer scheme.

e. Land surface processes

Treatment of land surfaces is a serious problem for atmosphere models. Surface fluxes of momentum, heat, and moisture are highly dependent on the land surface type and its horizontal homogeneity (Garratt 1993; Beljaars and Holtslag 1984). Almost invariably the surface characteristics vary on a scale smaller than the model's resolution. The current trend in atmosphere models is toward quite complicated treatments of the land surface (e.g., Sellers et al. 1986). In this paper we avoid most of these issues. We feel that a reasonable land surface parameterization needs to include the effects of topography, knowledge of the radiative heat balance at the surface, information on the ground hydrology and surface characteristics, and possibly more. These are problems that are beyond the scope of our current work, which is still primarily focused on atmosphere–ocean interaction.

We model the surface fluxes of latent and sensible heat as

$$Q_L = \rho C_Q L\mathbf{u}(q_{obs} - q), \quad (14)$$

$$Q_H = \rho C_Q C_p \mathbf{u}(T_{sfc} - T). \quad (15)$$

Here ρ is the air density, C_Q is an exchange coefficient, $\mathbf{u}$ is the lowest model level wind speed, q_{obs} is the observed 1000-mb humidity, q is the lowest model level humidity, T_{sfc} is the observed surface air temperature, T is the lowest model level temperature, and L and C_p have their usual meanings. Hence, we effectively relax the model T and q over land to observed values. [It should be noted that this does, however, prevent the model from attaining the right combination of humidity and moisture flux because, according to (14), the flux goes to zero if $q = q_{obs}$. This is clearly not right but it is difficult to know what else to do. The problem does not arise for temperature because T is typically less than T_{sfc}.)

While this is justifiable as an interim step, we will ultimately have to replace this formulation with a land surface parameterization that calculates T_{sfc} and the surface humidity. The former requires a reliable representation of the surface radiation budget that, because of our use of imposed radiative cooling profiles, is outside the scope of our model in its current form. Calculation of the surface humidity requires some representation of the ground hydrology, which again is beyond the scope of our current efforts. Alternatives, such as using the saturation humidity to calculate the potential evaporation and then reducing this by an empirically derived evaporation efficiency (e.g., Manabe 1969; Seager 1991), were found to lead to excessive evaporation.

The exchange coefficient used to calculate these fluxes was derived from the Deardorff PBL scheme as over the ocean. The roughness length for momentum over land was taken to be 0.1 m in general but to be 10 m over land with observed surface topography higher than 2000 m. The roughness length for heat and moisture was taken to equal that for momentum. The higher roughness lengths over mountains are designed to account for the effects of the great surface irregularity in those regions. It should be remembered that topography is otherwise ignored in the model.

f. Other model details

Several workers have noted that the free atmosphere appears to be damped on a quite short timescale (e.g., Holton and Colton 1972). Holton and Colton argued that the damping originated with cumulus momentum mixing, but this has been disputed. For example, Sardeshmukh and Held (1984) argue that the apparent damping is due to nonlinearities. Since nonlinear dynamics are neglected in the model, we parameterize their effect in a very simple way. First we include a vertical momentum diffusion, $\underline{F}$, of the form

$$\underline{F} = \epsilon \frac{\partial}{\partial\sigma}\left(\sigma \frac{\partial \underline{u}}{\partial\sigma}\right). \quad (16)$$

This form reduces the diffusivity upward in accord with expectations. Term ϵ is set so that the boundary-layer momentum diffusion would damp the flow on a timescale of 1.5 days. In addition, we use a Rayleigh friction damping with a timescale of 10 days so that, in sum, the lower atmosphere is damped on the shorter timescale and the upper atmosphere on the longer timescale.

Moisture is vertically diffused with the same formulation but with a timescale of only 5 days for the

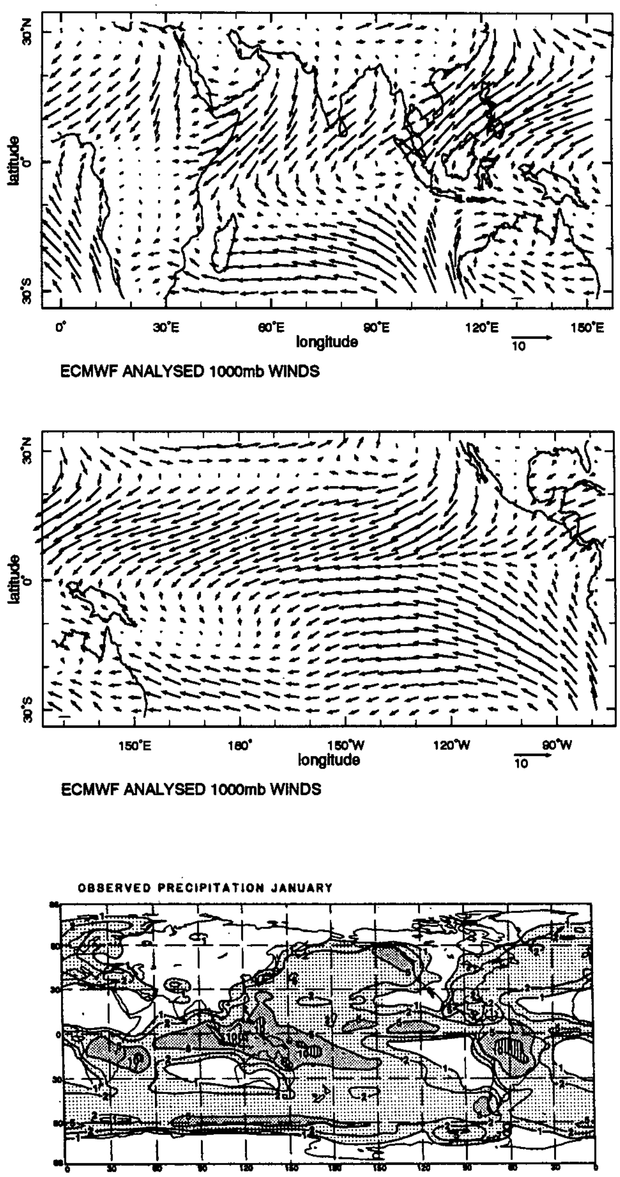


FIG. 1. Observed 1000-mb flow for (a) the Indian Ocean and (b) the Pacific Ocean sectors and (c) the observed precipitation (after Hart et al. 1990) for January.

lower atmosphere. Nonlinear vertical advections of temperature and humidity are included. This was done to ensure that the advected profiles were the near-neutral ones established by deep convection rather than the conditionally unstable basic-state profile. Advection is evaluated in physical space with an upwind difference scheme. No other nonlinear advections are included at present.

3. Model simulations

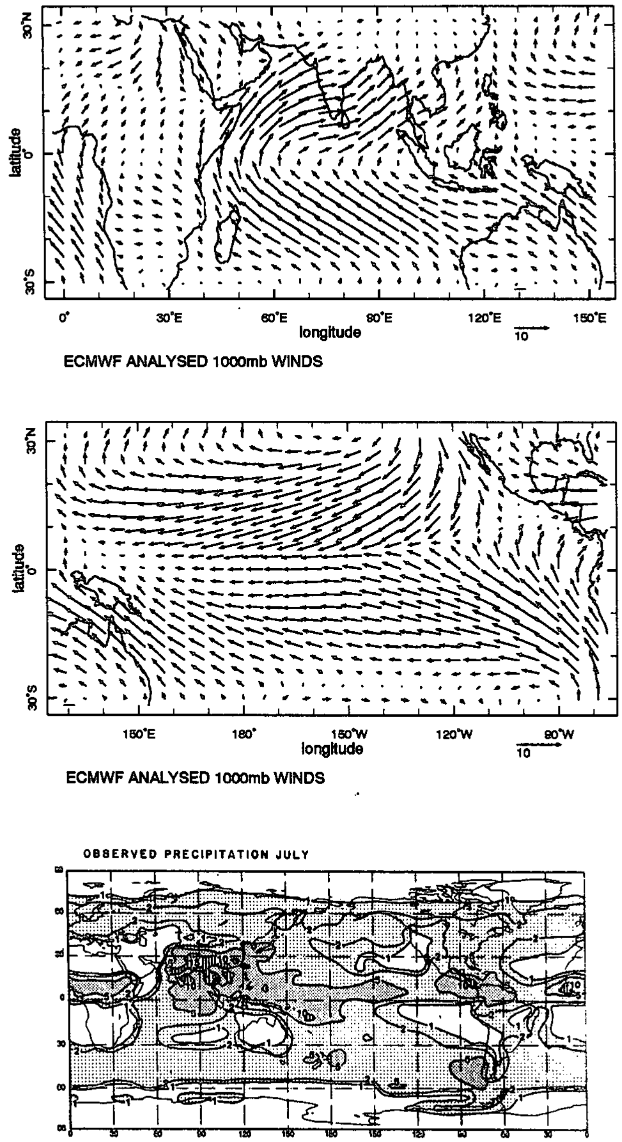
The atmosphere model was initialized at a state of rest with a temperature that was a function of the vertical coordinate only. The surface temperature was taken from the ECMWF analyses and represents monthly means for the period 1985–1992. The boundary conditions are the surface temperature over ocean and the surface air temperature and the 1000-mb humidity over land, all of which appear in the surface-flux calculations. The model time step is 1 day and the boundary conditions were linearly interpolated in time between the monthly means to derive daily values. For our first experiments the model was integrated through eight seasonal cycles. It was discovered that the model seasonal cycle essentially repeated itself after 2 years so subsequent integrations were for 4 years only. (To illustrate the degree of computational efficiency, a 4-yr integration uses about 3.5 h of CPU on a Silicon Graphics INDIGO^2 workstation.) All results are monthly means for the fourth year. [It is worth noting that the monthly means obtained from a seasonal cycle were quite different from those obtained when the model was run in perpetual-month mode. This has also been reported by Zwiers and Boer (1987).]

a. Sensitivity to convective parameterization

Here we present the tropical climate obtained from the model and the two versions of the Betts–Miller convection scheme. The model uses the combined Deardorff–Troehn and Mahrt PBL model and the simple cloud-radiation parameterization. The model precipitation and the 950-mb wind field are used as basic quantities with which to compare the performance of different model configurations. The first is important in its own right but is also a proxy for the distribution of latent heating in the atmosphere, which is an essential component of the forcing that drives the atmospheric circulation. It is also notoriously difficult to get right. The low-level wind field is presented because of its importance in driving the ocean circulation, and, ultimately, this model is intended for coupling to ocean models. It is more common in the literature for atmosphere models to be verified against upper-tropospheric quantities and sometimes the 850-mb winds. Our focus on air–sea interaction dictates that we must instead verify the model against near surface quantities.

Figures 1 and 2 show the observed 1000-mb winds and precipitation in millimeters per day for January and July taken, respectively, from the ECMWF analyses and from the climatology of Jaeger (1976). Figures 3 and 4 show the same quantities from the fourth year of a seasonal-cycle integration of the model with the BM86 convection scheme. The lack of nonlinear dynamics in the model and the tapering away with latitude of the radiative cooling means that the circulation is very weak outside of the Tropics and the precipitation is weak as well. We restrict comparison to the regions equatorward of about 35°N and 35°S.

During January the model with the BM86 scheme does capture some elements of the observed distribu-



tion of the precipitation, including heavy rain to the south of the equator over South America, the large region of rain near to the equator over the Indian Ocean, the marked South Pacific convergence zone (SPCZ), and the intertropical convergence zone (ITCZ) in the east Pacific Ocean. It misses the rainfall over tropical Africa, Indonesia, and the Atlantic ITCZ. Further, the rainfall in the east Pacific ITCZ is too weak, although the SPCZ, Indian Ocean, and South American rainfall are about correct. The SPCZ extends too far east, which is a common problem in atmosphere models, including sophisticated GCMs.

The model does a better job over Africa in July in both pattern and magnitude but now underestimates the rain over Central America. The east Pacific ITCZ remains much too weak and, as in all months, the SPCZ extends too far east. An even more obvious error is the totally inadequate simulation of the Asian summer monsoon rainfall. The heavy rain over the Indian Ocean to the north of the equator is simulated, but the model fails to produce heavy rain over the continental areas of Southeast Asia. This is perhaps surprising in that the land-surface flux scheme relaxes the low-level temperature and humidity to those observed, and they have particularly high values in this region. Clearly the model is excluding some important process here. In all

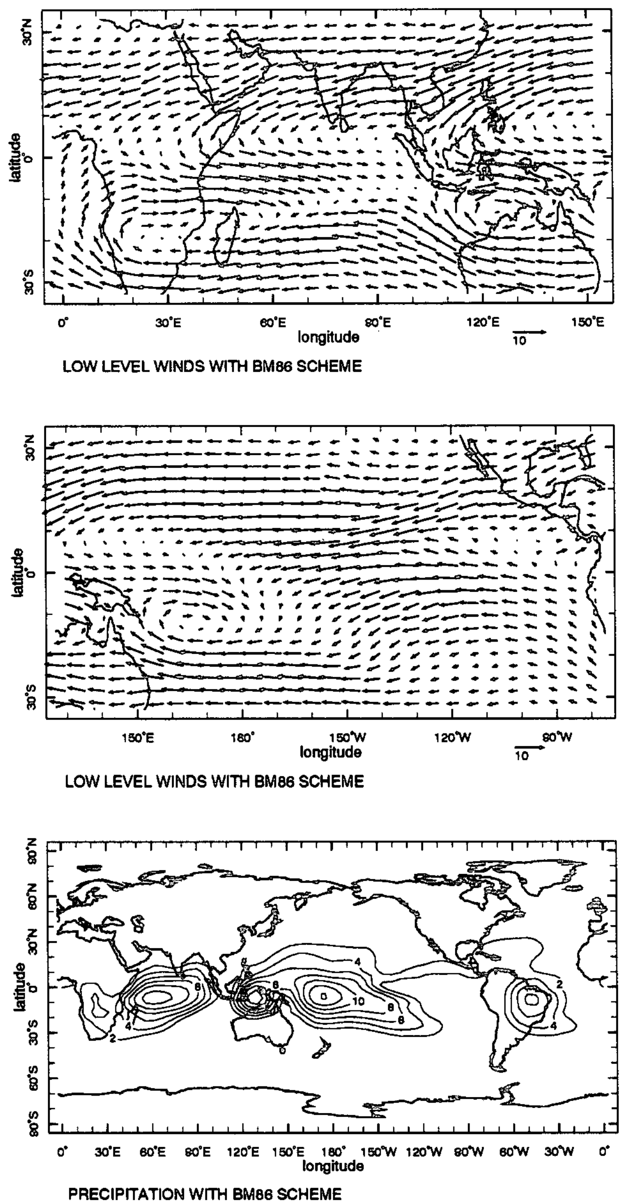


Fig. 3. Low-level wind over (a) the Indian Ocean sector and (b) the Pacific Ocean sector and (c) the modeled precipitation for January and the case with the Betts–Miller (1986) convection scheme.

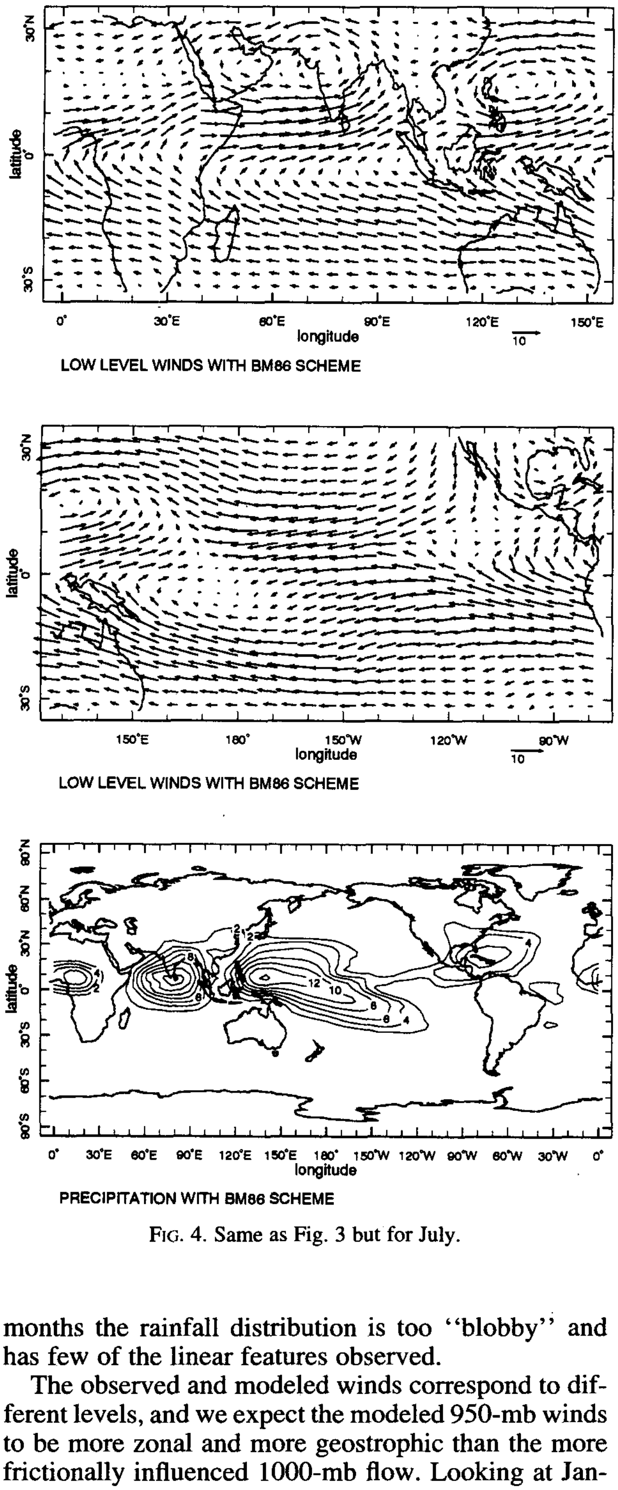


FIG. 4. Same as Fig. 3 but for July.

months the rainfall distribution is too "blobby" and has few of the linear features observed.

The observed and modeled winds correspond to different levels, and we expect the modeled 950-mb winds to be more zonal and more geostrophic than the more frictionally influenced 1000-mb flow. Looking at January first (Figs. 1 and 3), in the Indian Ocean sector the most obvious errors involve westerlies on the equator around 60° and 130°E, which are far stronger than observed. However, the northeasterly flow over north Africa and all of the Asian continent is reproduced but, as expected, is more zonal than the observed 1000-mb circulation. The southeasterly flow south of the equator is also reproduced. In the Pacific sector the flow is reasonably well reproduced, although, again as expected, it is more zonal than the 1000-mb flow. The equatorward sweep of the trades into the convergence zones is apparent. However, off the coasts of Mexico and South America the flow is far too zonal, even allowing for possible differences between the 1000- and 950-mb levels. Indeed the model produces easterly flow at 80°W south of the equator where the observed winds are purely southerly. The same problem occurs off the Mexican coast. Excessive zonality could be related to the weak ITCZ or the lack of topography that blocks east–west flow. In the west Pacific the model again overestimates the equatorial westerlies.

During July (Figs. 2 and 4) the most obvious model error is related to the weak Asian monsoon. The model does simulate southeasterly and southwesterly winds over the Indian Ocean that feed the monsoon, but the absence of convection in the region around 110°E and 20°N means a cyclonic circulation develops in response to heating to the east. In the observations, the flow in this region is dominated by the convergence of the Pacific trade winds into the monsoon trough and, hence, a weak southeasterly flow. The model errors here are clearly related to the errors in the precipitation field. Over the Pacific sector the model does simulate the general sense of the trade winds and convergence zones. The main errors are again the zonality of the flow in the east Pacific and south of the equator and the development of westerly flow in the western equatorial Pacific. This latter feature of the modeled circulation means the equatorial easterlies do not extend far enough west and instead dip southward across the equator into the SPCZ, which, as we have seen, is too strong in its eastern extension.

Figures 5 and 6 show the same quantities obtained by a model integration with the BM93 convection scheme. All other model parameters are set the same, with two exceptions. Lower values of the subsaturation are assumed with the BM93 scheme, together with a slightly different clear-sky radiative cooling profile. Both were chosen, within reason, to give the best results with each convection scheme. In both January and July, marked improvements in the precipitation simulation occur. In January the Atlantic and east Pacific ITCZs are strengthened, the pattern of rainfall over South America is improved (although it underestimates it now), a rainfall maximum appears over Africa, and the SPCZ is more confined to the west Pacific as observed. The July precipitation simulation is also improved in a similar way with the SPCZ more restricted to the west, a reduction of excess rain over the Indian Ocean, a stronger east Pacific ITCZ, and stronger rain over the Indonesia area. In both months the areas of convection are more meridionally restricted, with a particularly marked improvement in the west Pacific. The precipitation pattern obtained with the revised Betts–

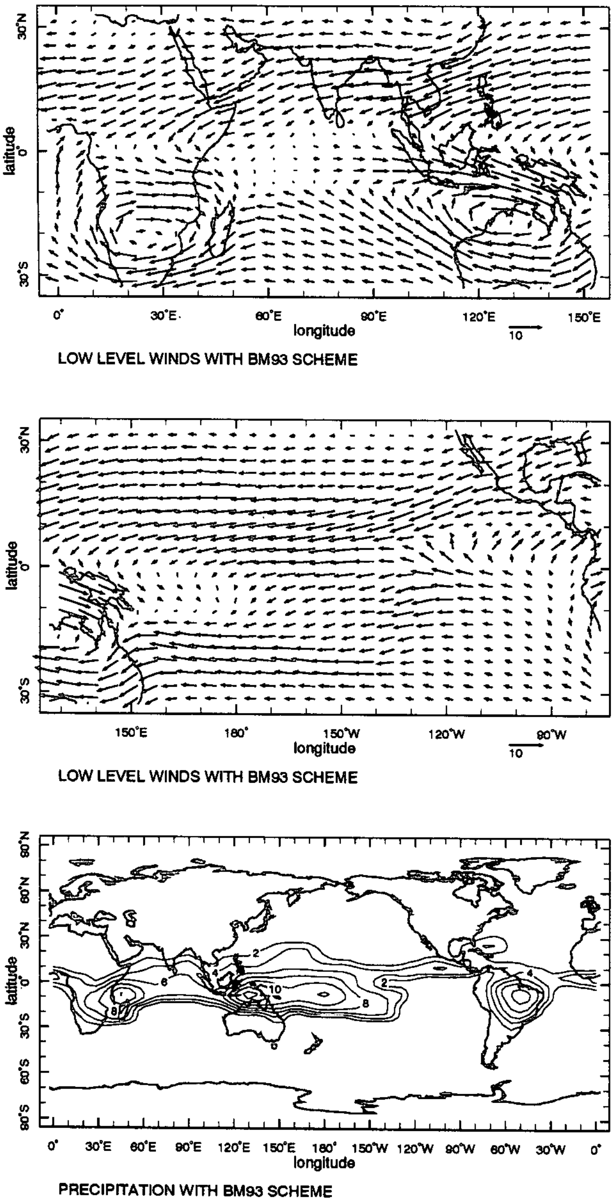


FIG. 5. Low-level wind over (a) the Indian Ocean sector and (b) the Pacific Ocean sector and (c) the modeled precipitation for January and the case with the Betts–Miller (1993) convection scheme.

Miller scheme is more linear than with the original scheme and qualifies as a major improvement.

The low-level winds obtained with the BM93 scheme are also, in some ways, improved over the BM86 case. In particular the southerly flow across the equator in the east Pacific during northern summer is much better simulated. While this is a clear response to increased convection in the ITCZ, the otherwise modest changes in the wind field indicate the extent to which low-level tropical flows are determined by processes other than forcing associated with convective heating. Further, in January, despite an improvement in the simulation of the ITCZ rainfall in the BM93 case, the southeasterly flow south of the equator is actually *less* well represented than in the BM86 case. Clearly, simulation of low-level flow depends on far more than just getting the distribution of convective heating correct. However, the general improvement of the BM93 model over the BM86 model led us to adopt it as the "standard" model for which we will describe the model climate.

In Figs. 7 and 8 we show the observed and modeled zonal flow on the equator as a function of height and longitude (the Walker circulation) for January and

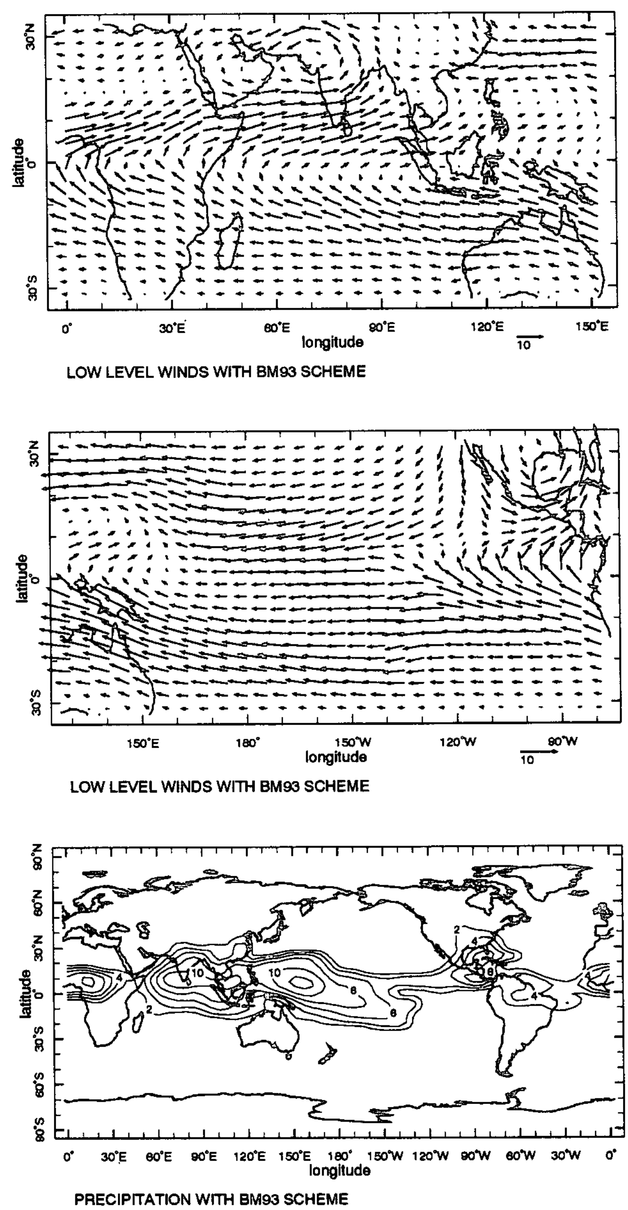


FIG. 6. Same as Fig. 5 but for July.

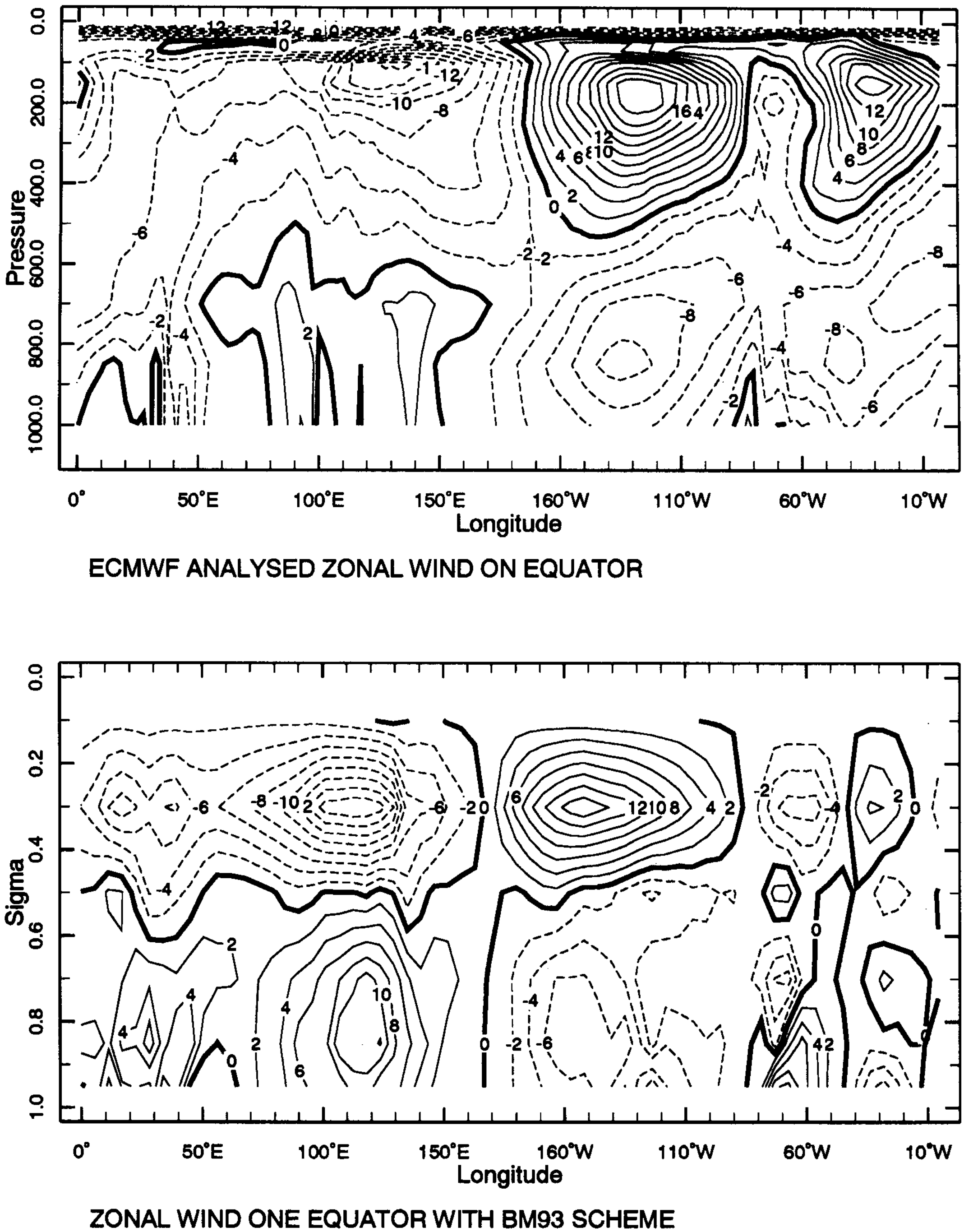


FIG. 7. (a) Observed and (b) modeled zonal flow on the equator as a function of height and longitude for January for the case with the BM93 convection scheme.

July. The January comparison is quite favorable with the strong overturning over Pacific and Indian Ocean sectors captured quite well. The exception is that the westerly winds centered at 120°E are too strong. This reveals the expected model westerly bias caused by exclusion of nonlinear advections that prevents zonal mean easterlies on the equator. The model also fails to simulate the strength of the upper-level westerlies over South America and places all the upper-level flow maxima too low (in part a vertical resolution problem). Problems with the upper-level flow are more acute in July. Upper-level westerlies are produced over the Pacific where weak easterlies are observed in most places. This is most likely another consequence of the poor

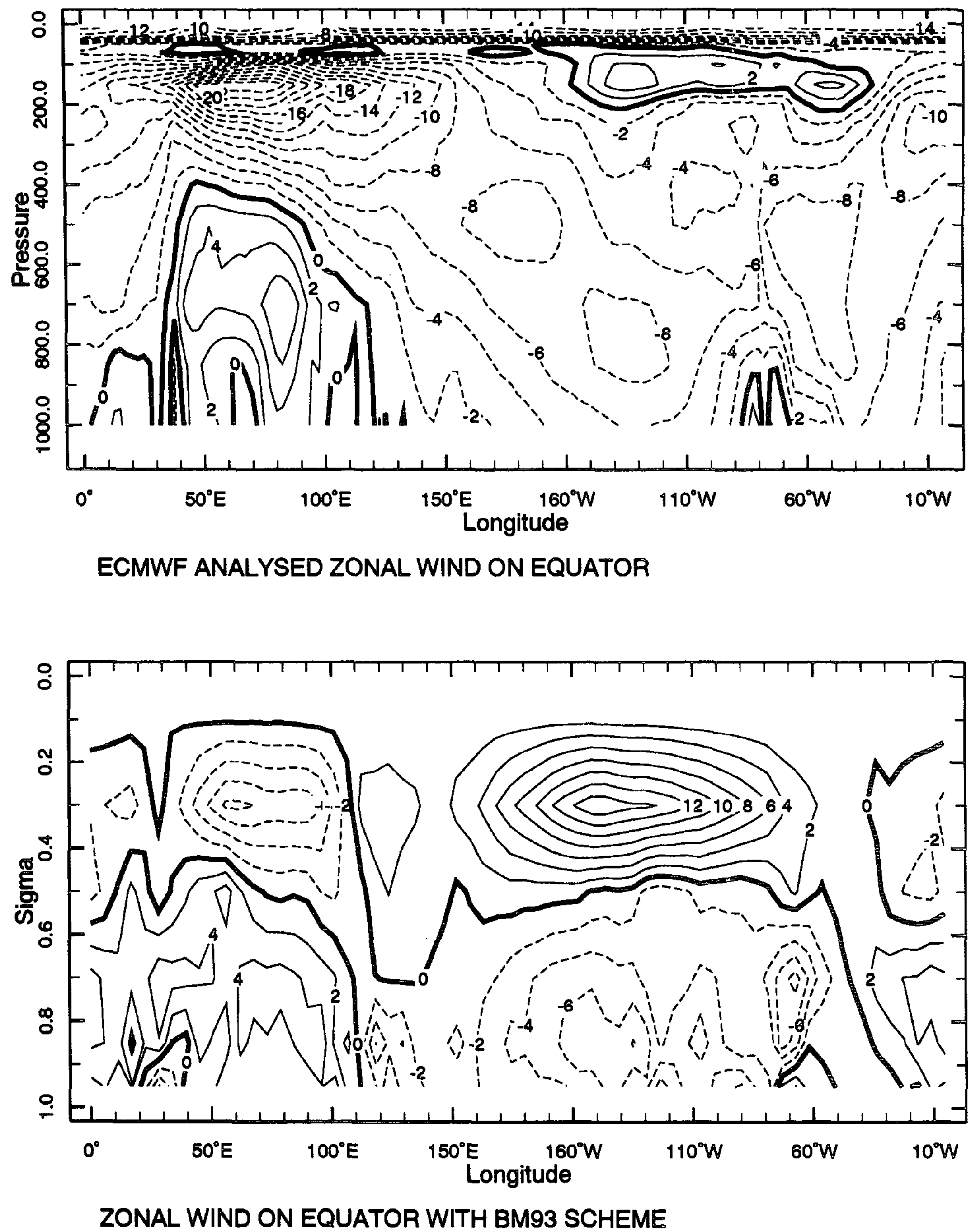


FIG. 8. Same as Fig. 7 but for July.

summer monsoon in the model. In nature the monsoon draws convection north and off the equator over the west Pacific. In the model extensive convection remains on the equator, which forces Kelvin waves and upper-level westerlies to the east that are not excited by the observed off-equatorial convection.

In contrast, the low-level flow in July is captured quite well. This contributes to one of our general conclusions: reliable simulation of the quantities most relevant to air–sea interaction does not critically depend on equally accurate simulations of the entire troposphere. The errors seen in July would, however, likely impact on modeled teleconnections that flow through the upper troposphere and does need to be addressed (Webster and Chang 1988). Nonetheless, the large difference in quality of simulation of the upper and lower troposphere does raise the question of the extent to which anomalous circulations at upper levels can influence the lower atmosphere.

The seasonal cycles of lowest-level zonal and meridional winds on the equator are shown in Figs. 9 and 10. The simulation is generally reasonable. The model has a strong seasonal cycle in cross-equatorial flow in

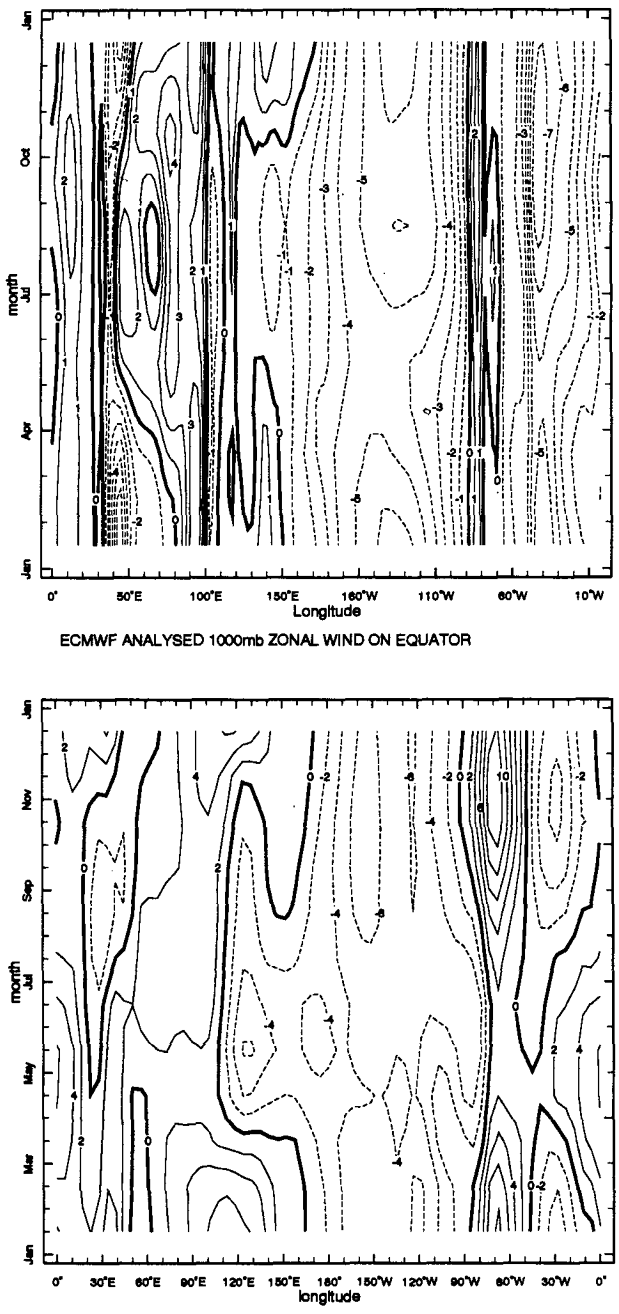


FIG. 9. (a) Observed and (b) modeled zonal flow on the equator as a function of longitude and time for the case with the BM93 convection scheme.

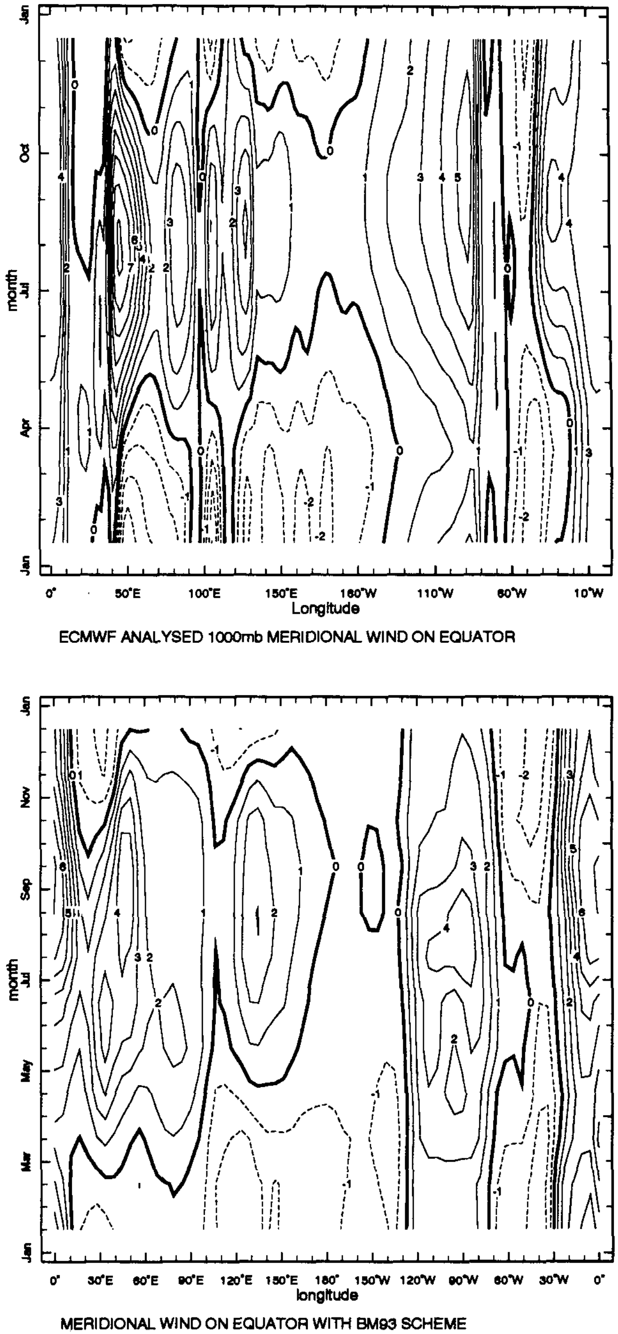


FIG. 10. Same as Fig. 7 but for meridional flow.

the east Pacific, although it peaks too early in the summer. The seasonal cycle of easterlies over the Pacific is well represented in amplitude and phase. Strong seasonal cycles in cross-equatorial flow are also seen over the Atlantic and Indian Oceans and Africa and are modeled reasonably. As expected, the model exhibits a westerly bias compared to observations.

The surface stress over the Pacific is shown in Fig. 11 for January and July. These can be compared with,

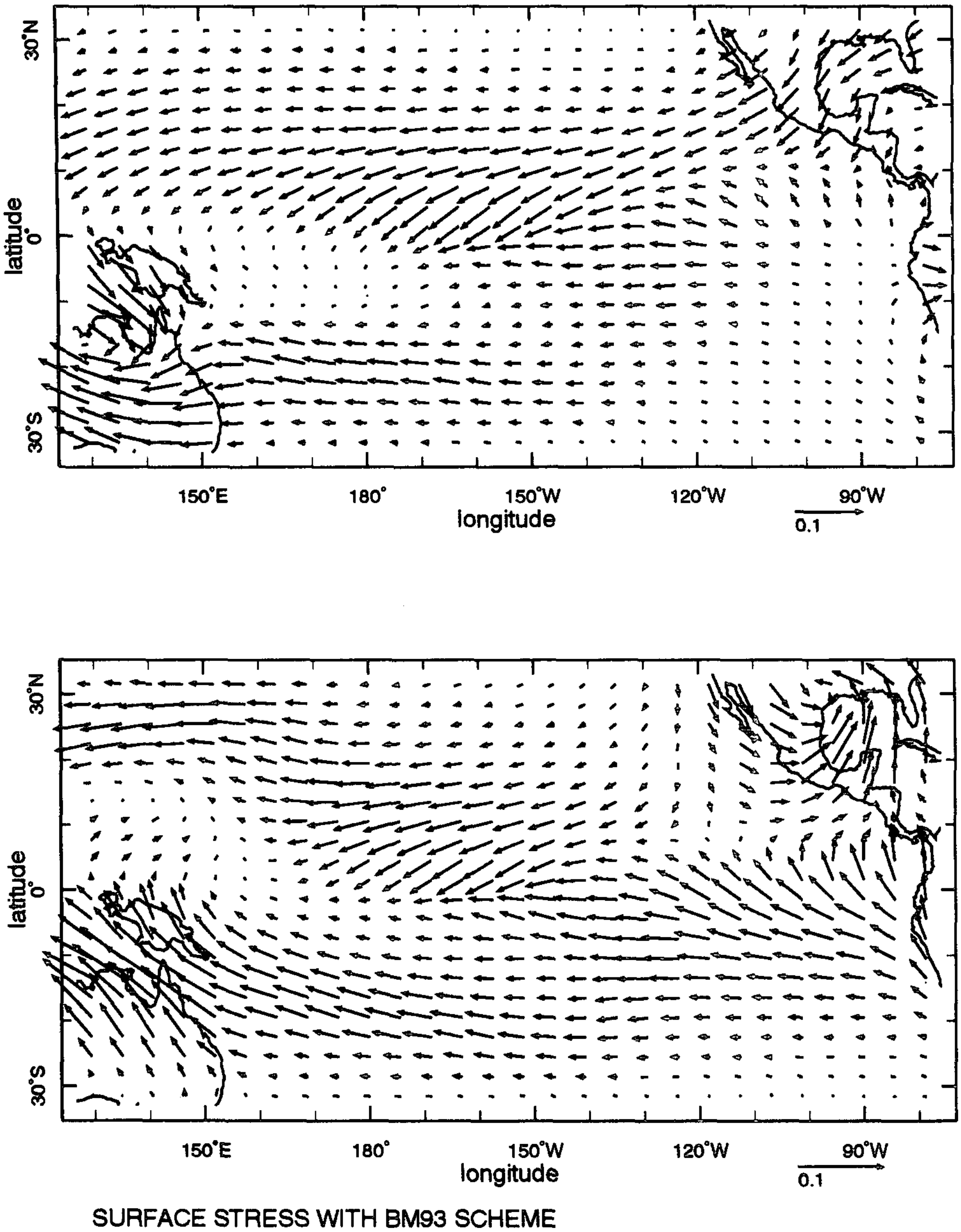


FIG. 11. Modeled surface stress (kg m^{-1} s^{-2}) over the tropical Pacific for (a) January and (b) July for the case with the BM93 convection scheme.

for example, the stresses derived from observations by Hellerman and Rosenstein (1983, not shown). Clearly the surface stress is more meridional than the low-level wind field. In January the northeasterly stress north of the equator is quite well captured. The major model failing is an underestimate of the stress in the southeastern Pacific, which is a result of the weak winds. In July the stress in the southeast Pacific is quite well represented, including the northward stress right on the equator. The strong stress northeast of Australia is well represented, as is the stress under the northeast trades. However, in both months the location of the maximum zonal stress is located too far equatorward.

In summary the model produces a tropical climate, as defined by precipitation and low-level winds, that is, in some important respects realistic despite some systematic errors. The modeled precipitation is far more realistic with the revised convection scheme, which accounts for the effects of convective downdrafts.

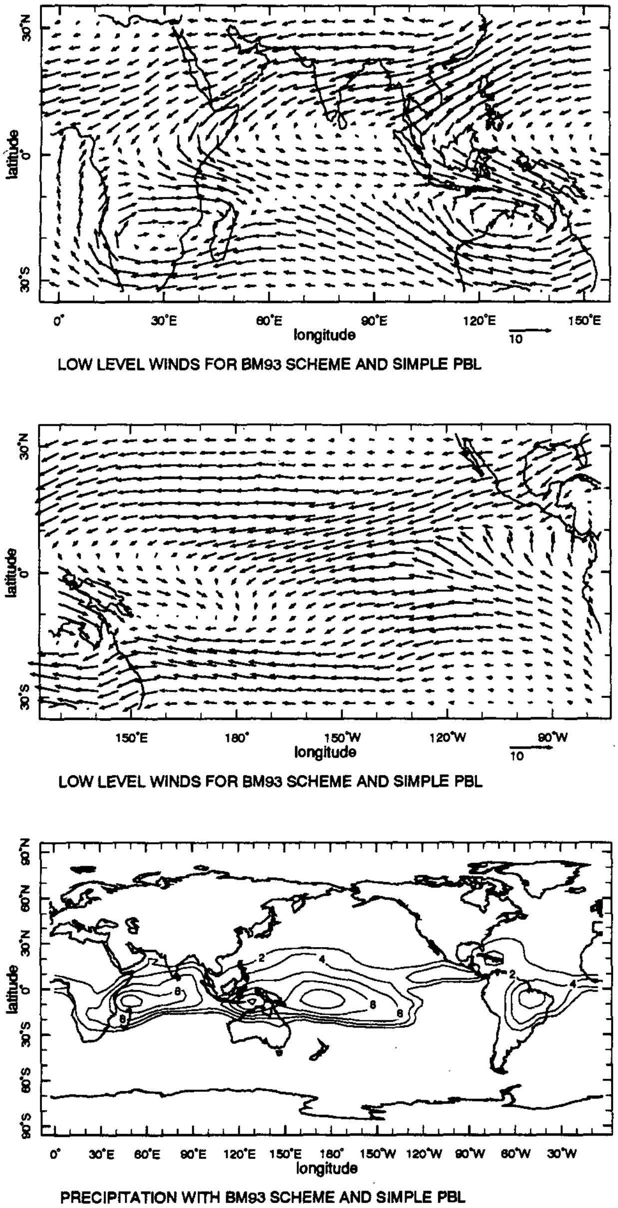


FIG. 12. Low-level wind over (a) the Indian Ocean sector and (b) the Pacific Ocean sector and (c) the modeled precipitation for January and the case with the BM93 convection scheme and simplified PBL.

b. Sensitivity to PBL physics

We completed a seasonal cycle integration of the model with the combined Deardorff–Troehn and Mahrt PBL scheme removed. In its place we used constant exchange coefficients of 0.0015 for heat and moisture and 0.0012 for momentum over ocean and twice these values over land. If the virtual potential temperature of the 950-mb air exceeded that of the surface by more than 3 K, then the exchange coefficients were reduced to zero for heat and 0.0005 for momentum to take account of the low-level stability. The PBL was assumed to have a top at $\sigma = 0.85$ and was mixed toward uniform profiles of potential temperature and humidity with a timescale of 4 days.

The precipitation and $\sigma = 0.95$ winds are shown in Figs. 12 and 13. For January the main changes are a modest weakening of the oceanic convergence zones and a larger reduction of rainfall over the tropical continents that becomes too low. Apparently the full PBL scheme is capable of generating large exchange coefficients over land that aid the precipitation simulation through the effect on the local evaporation. Removal of the PBL scheme leads to only very slight changes in the flow field. The only exception is that during Jan-

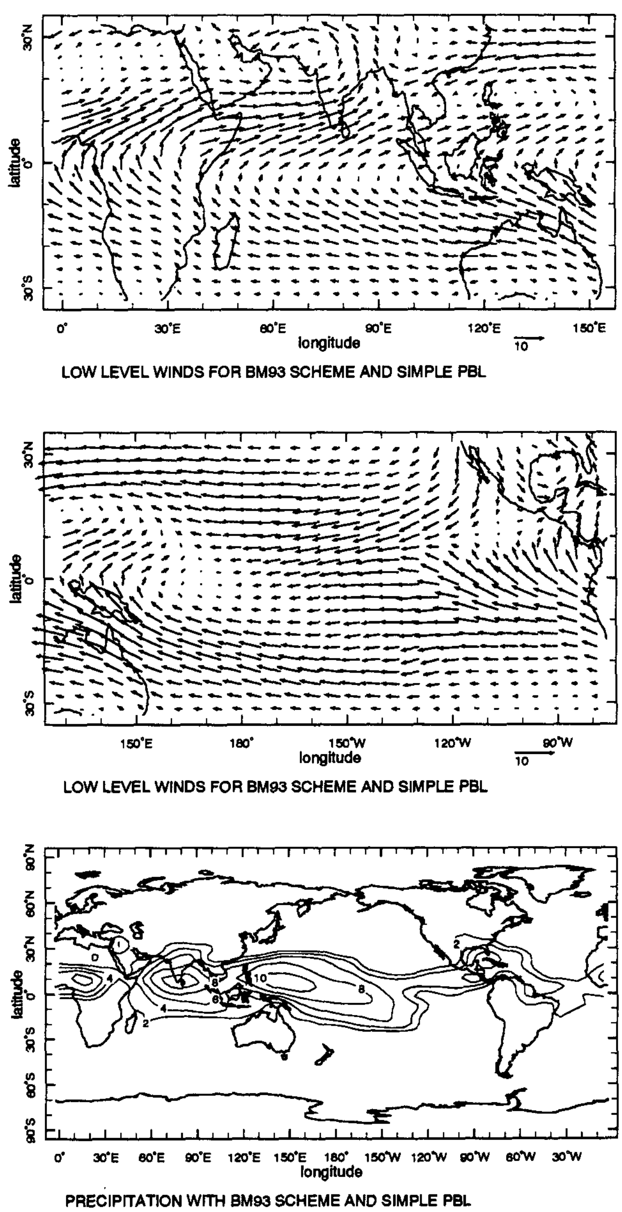


FIG. 13. Same as Fig. 12 but for July.

uary the southeasterly flow in the southeast equatorial Pacific is more faithfully represented than in the case with the full PBL physics. This suggests that the poor simulation of this flow in the standard case is the result of errors in the treatment of surface friction introduced by the PBL model.

During July removal of the PBL again has relatively modest effects on the precipitation but does restrict the rainfall over the Indian Ocean sector even more to the ocean regions. Rainfall is again slightly reduced over the tropical land areas. The changes in the low-level wind field are very small. However, the westerlies in the west Pacific are strengthened, which follows from the weaker Asian monsoon. Reduction of rainfall over Central America greatly reduces the strength of a related largely spurious cyclonic circulation. Since the real atmosphere in this region is strongly effected by topography, which blocks east–west flow, this improvement is largely fortuitous and is gained at the expense of the precipitation simulation.

Overall the differences in the model simulation between the case with a detailed PBL parameterization and one with only a very simple scheme are small. The one major difference is in the direction of the surface stress (not shown), which in the case without the PBL scheme, must be in the direction of the lowest model layer wind but, with the PBL scheme, is directed considerably more toward lower pressure. The latter is more realistic. However, the similarity of the modeled low-level winds indicates this realism is of more importance for the forcing of ocean motions than it is for the atmosphere. We will return to why the model is relatively insensitive to a major change in model physics in section 5.

c. *Sensitivity to radiative cooling*

Consideration of simple models of a convecting atmosphere suggests that the tropical climate should be strongly controlled by radiative processes. Several authors have shown that the amount of evaporation, the largest energy source for the atmosphere over the ocean, is given by the amount of radiative cooling, the only energy sink for the atmosphere other than friction (e.g. Sarachik 1978; Betts and Ridgway 1988). Since, integrated in space, the precipitation must balance the evaporation, we expect the amount of precipitation, and hence the strength of convective heating, to be strongly determined by the radiative cooling.

We completed a model integration in which the imposed radiative cooling was halved everywhere. From the foregoing argument we would expect the rainfall to be lower by a similar amount. In fact it was lowered by about one-third, with the discrepancy probably arising from neglect of moisture transport by boundary-layer mixing and diffusion in the theoretical argument. Nonetheless, this results shows the extent to which the precipitation is dependent on the parameterization of radiation. Since, in this model, the radiative cooling is externally imposed, the total amount of precipitation is quite arbitrary; it would become significant only in the presence of a more realistic treatment of radiation.

A number of other experiments were carried out in which the imposed radiative cooling profile was altered, all of which classify as examples of what we found did not work. For example, assuming a profile with uniform cooling below $\sigma = 0.3$ led to excessively large areas of convection. This demonstrates how the vertical profile of radiative cooling, and in particular the decrease of cooling with height, controls the degree of convective instability. Removing the cloud–radiation parameterization, while not having much effect on the distribution of rainfall, led to deep convective heating profiles that peaked too low in the atmosphere. This seriously degraded the low-level wind field, indicating the importance of the vertical profile of diabatic heating.

An interesting experiment that also illustrates a model configuration that does not work replaced the imposed cooling profile with a clear-sky radiative transfer calculation. For simplicity we only account for water vapor absorption, which dominates the heating and cooling in the solar and longwave spectrums. For longwave radiation we used the nonisothermal emissivity formulation of Ramanathan and Downey (1986), which has been shown to be fast and accurate (and, like the convection scheme, was coded in house and makes extensive use of look-up tables to increase computational speed). For solar radiation we used the formulation of Lacis and Hansen (1974).

Figure 14 shows the rainfall for this case. Deep convection was very widespread, almost ubiquitious, throughout the Tropics. Shallow convection was absent, which appears to be the key to understanding this result. The absence of cloud-top radiative cooling prevented the destabilization that generates and maintains shallow convection. This allows conditions favorable for deep convection to develop. In the standard model we force shallow destabilization to occur by adopting a profile with maximum cooling at $\sigma = 0.85$ and then imposing the cloud–radiation feedback. With this profile shallow convection occurs in regions without deep convection, drying the lowest level and moistening the layer above. The moistening above is balanced by subsidence drying. If, with clear-sky radiative transfer and no cloud-radiation interaction, this cannot occur, then the model is forced to balance the surface flux with deep convective drying. It has no choice about this because the vertical velocity in the lowest level is always insufficient to balance the surface flux with subsidence drying. Slingo and Slingo (1988) also present a GCM simulation with no cloud–radiation interactions that showed broadened regions of tropical rainfall.

This last experiment illustrates the importance of cloud radiation interaction in regulating tropical convection. Dynamics alone appear insufficient to explain

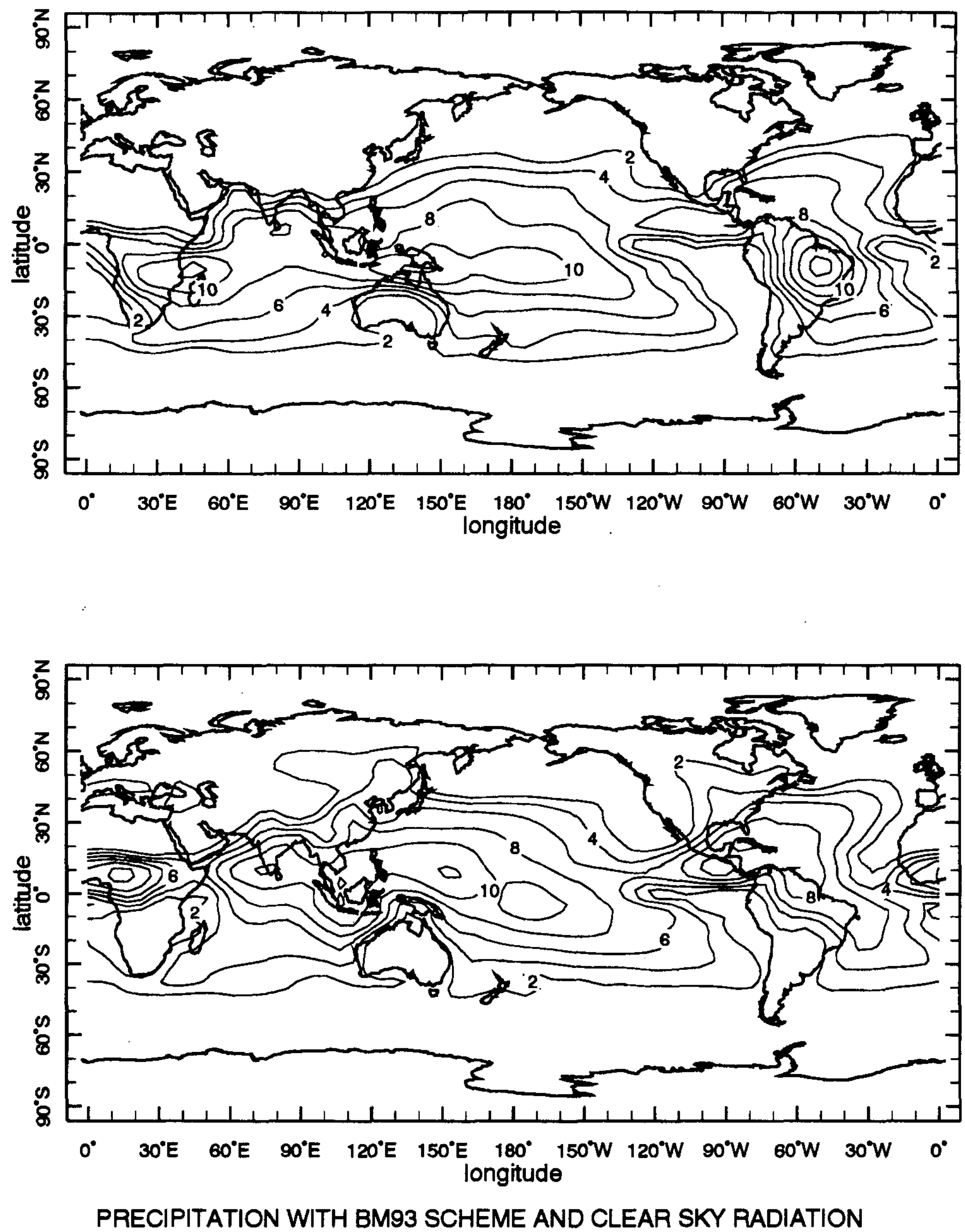


FIG. 14. Modeled precipitation for (a) January and (b) July for case with BM93 convection scheme and clear-sky radiative transfer.

the scale organization of tropical convection. More generally, all these experiments illustrate the extent to which radiative processes control the extent and vertical distribution of convective heating.

4. Experiments with an imposed monsoonal heat source

As has been noted, the model's representation of the Asian summer monsoon is very poor. We believe this is due to the lack of topography in the model. Many investigators and modelers, dating back to the early experiment of Manabe and Hahn (1975), have demonstrated the role that the mountains of Asia play in determining the location and intensity of monsoon rainfall. Although the mountains dynamically effect the large-scale stationary wave pattern, the effect most argued for is the generation of orographic rainfall.

Work is underway to include mountains in the model. For now we are interested in the extent to which a poor monsoon is degrading the model simulation in other regions, particularly over the Pacific Ocean. To address this we impose a heating centered over Southeast Asia and with a horizontal scale comparable to that observed. The heating peaks at 500 mb and is approximately sinusoidal in pressure. The heating grows in time from May to a maximum intensity in August (7 K day^{-1} at 500 mb), then decays and is zero during northern winter. It thus approximates the spatial and temporal pattern of the observed Asian monsoon.

Figure 15 shows the surface stress over the Pacific for July of a seasonal cycle integration with this imposed heating. Comparing this with the standard model the changes are seen to be relatively modest. The trades over the Pacific sweep a little farther west and then northwestward into the monsoon trough, which is slightly more realistic. Over the Indian Ocean (not shown) there are stronger southeasterlies north of the equator. Kanamitsu et al. (1990) have also shown that a poorer simulation of the Asian monsoon (in their case this was the result of reducing the resolution of the National Meteorological Center model) degrades the simulation of the wind field over the west Pacific, but the effect in their model was much larger than seen here.

Our results suggest that improving the models monsoon, and by implication inclusion of orography, are priorities not only in their own right: the large-scale circulation will also be affected. For the tropical Pacific the effects are modest, suggesting that the surface winds are more determined by the local convection than that over Asia. It remains to be demonstrated just how closely the *variability* of the Asian and Pacific sectors are coupled.

5. Discussion

We have presented aspects of the tropical climate simulated by a linear primitive equation model and a minimal package of physics parameterizations. We have attempted to show how the modeled climate depends on assumptions regarding the treatment of convection, radiation, and boundary-layer processes.

It was found that the model climate was quite different depending on whether the early version or the revised version of the Betts–Miller deep convective adjustment scheme was used. The distribution of precipitation and the low-level wind field was unambiguously more realistic with the revised version. The former scheme adjusts the atmosphere, in regions of gravitational instability, to a profile drawn through the low-level θ_E and then corrects this profile to ensure conservation of moist static enthalpy. The correction typically ensures drying and cooling of the lowest model level. However, this drying and cooling is accomplished in a rather ad hoc manner theoretically removed from the activity of convective downdrafts that are actually responsible.

In the latter scheme the reference profile in the lowest levels is instead given by the forced descent of unsaturated environmental air from above the boundary layer. This introduces colder and drier air into the boundary layer in a manner qualitively similar to that observed. This very simple change has a dramatic effect on the simulated precipitation. With the earlier scheme the precipitation was characterized by a "blobby" distribution featuring large regions of precipitation in excess of many observations over tropical Africa, the Indian Ocean, and the west Pacific. Linear features such as the ITCZs were weak, the SPCZ extended far too far east, and the general meridional extent of precipitation was too large.

The downdraft parameterization reduces many of these problems. The precipitation pattern becomes more linear with the rainfall reduced in regions of maxima but increased in the ITCZs. The meridional extent of rainfall is reduced and the SPCZ does not extend as far east (though it still extends too far). These changes are physically reasonable and understandable in terms of the effects of downdrafts. Downdrafts have a strong stabilizing effect by injecting low entropy air from above into the boundary layer. This acts to oppose the destabilizing effects of surface fluxes and moisture convergence and strongly regulates convection. The regulation prevents the development of "blobs" with excessive rain and redistributes the convection to the regions of high low-level θ_E air, which typically lie above the warmest water.

While these changes are a major improvement the rainfall distribution remains too broad meridionally. In particular, if we believe the satellite rainfall data recently acquired from the Microwave Sounding Unit (MSU) by Spencer (1993), then the modeled rainfall is quite wrong. The MSU data shows the heaviest rain to be in the east Pacific ITCZ and the regions of precipitation to be much more linear than in our model. The model rainfall is more akin to the outgoing longwave radiation (OLR) rainfall data (e.g., Arkin and Janowiak 1991). This is true of almost all existing atmosphere models [e.g., Kanamitsu et al. (1990) at NMC; Arakawa and Cheng (1993) at UCLA; McFarlane et al. (1992) at the Canadian Climate Centre; Hart et al. (1990) at the Australian Bureau of Meteorology Research Center; Tiedtke et al. (1988) with the ECMWF model and a Kuo convection scheme; and Sud et al. (1992) and Moorthi et al. (1994) with different versions of the Goddard Laboratory for Atmospheres model, to cite just a few]. One exception to this perhaps discouraging rule is the United Kingdom Meteorological Office model, which uses a mass flux scheme (Gregory and Rowntree 1990). This model does underestimate the ITCZ rainfall compared to MSU data but simulates the SPCZ very well. The version of the ECMWF model with the BM93 scheme does produce a very strong, linear, ITCZ, which is the only model simulation qualitatively similar to the MSU patterns in this area (Betts and Miller 1993). However,

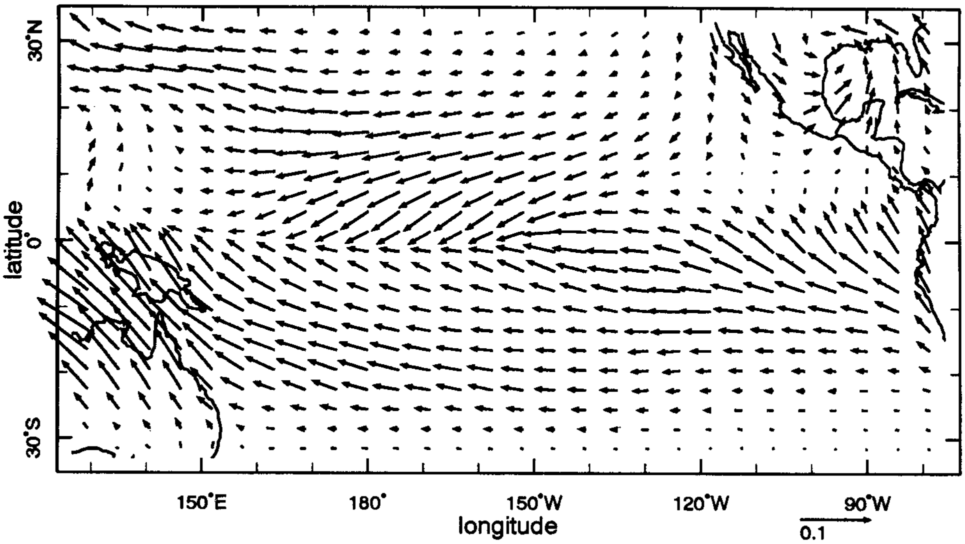


FIG. 15. Modeled surface stress (kg m^{-1} s^{-2}) in July over the Pacific Ocean for standard model with an imposed, seasonally varying monsoonal heat source over Asia.

it should be remembered that there is still doubt as to whether the MSU values are correct (Spencer 1993).

We found that the model precipitation magnitude, but not pattern, was very sensitive to the magnitude of the radiative cooling. This is expected in that the evaporation, and hence precipitation, must balance the radiative cooling (Sarachik 1978). The most interesting experiment used a clear-sky radiative transfer calculation and resulted in almost ubiquitous deep convection in the Tropics and an absence of shallow convection. This argues for the role of cloud–radiation interaction in maintaining regions of shallow convection and restricting the areas of deep convection. The strong dependence of the model on the radiative cooling underlines the need to advance to a physically based calculation of radiative cooling. We are working on combining the clear-sky radiative transfer scheme we already have with the cloud-radiation schemes of Kiehl et al. (1987) and Lacis and Hansen (1974) and the cloud-generation schemes of Slingo (1987) and Betts and Boers (1990).

The sensitivity to radiative cooling is in contrast to the low sensitivity of the model to the modeling of boundary-layer processes. Replacement of a detailed PBL scheme of GCM level complexity with one that assumes a constant PBL depth and constant exchange coefficients effected a quite modest change in the simulation. This is understandable in light of the radiation experiments. Given that the radiative cooling remained unchanged in the two PBL runs, the surface fluxes of sensible heat and moisture were constrained to be very similar. While the exchange coefficients were different, the model's low-level thermodynamic quantities adjusted so that the surface fluxes satisfied the necessary energy constraint. While this is not an unexpected result, it provides another justification for paying more attention to radiation. More generally the atmosphere is seen to be constrained by the assumptions regarding energy sinks. Given our crude assumptions in this respect, attempting to further improve the model simulation by tinkering with aspects of the internal thermodynamics is, at this stage, clearly a fruitless task.

Since this model was developed to overcome the limitations of simpler models such as that of Seager (1991), it is useful to see whether the greater complexity has rewarded us with an improved simulation. First, Seager (1991) found it necessary to include the observed low-level humidity as a model forcing because attempts to compute it were unsuccessful. This problem has been overcome here, which is a major improvement in generality. Major problems of the Seager (1991) model, which is also true of models with similar structure (e.g., Neelin 1988), were the weak meridional winds and underestimation of the strength of the ITCZs. The mass convergence in the east Pacific ITCZ of the current model is about three times stronger than in the earlier model. This is partly due to the strength of the convective heating in the ITCZ. However, strong mass convergence also occurs with the BM86 convection scheme even though the convective heating is much weaker in that case. In both cases it appears that the cross-equatorial flow is being partly driven by the establishment of pressure gradients near the surface by the strong SST gradients, in accordance with the ideas of Lindzen and Nigam (1987). The simple vertical structure adopted in the earlier model was incapable of capturing this process, although modifications of the Gill model have since tried (Wang and Li 1993).

The current model exhibits numerous other more detailed improvements. These are achieved without needing to specify unrealistically large friction throughout the atmosphere (instead strong friction is confined to the boundary layer), raising hopes that the model will be able to capture teleconnections that propagate through the upper troposphere once inclusion of nonlinear terms allows generation of a realistic mean meridional circulation.

6. Summary and conclusions

This paper has reported on progress in the development of a computationally efficient but realistic global atmosphere model for use in long-term climate studies. The goal is to develop a model that avoids the pitfalls and problems encountered with highly simplified models while avoiding the computational expense of a typical GCM. To this end we have developed and utilized a method of solving the primitive equations involving a complete normal-mode decomposition. This allows unconditionally stable analytical time integration and use of long time steps, resulting in the major gain in efficiency relative to more traditional models. We use a time step of one day. The model numerics and architecture have been described in Seager and Zebiak (1994).

Our purpose in incorporating physical parameterizations is to proceed step by step rather than to immediately adopt a full physics package from an existing GCM. There are two reasons for this. First, by including processes gradually we are able to assess the importance of each added physical parameterization. Second, we do not wish to sacrifice the computational efficiency of the model dynamics by adopting slow and unwieldy physics packages unless there is no alternative. To date we have included a convection scheme that is simple and fast but has also been demonstrated to be reliable (Betts and Miller 1986, 1993) and a PBL parameterization resting on the familiar similarity theory. Radiation is parameterized as an imposed cooling, and land surface processes are largely sidestepped by relaxing the model to observed surface temperature and near-surface humidity on timescales computed internally by the model. The model has a low vertical resolution of only six layers, and the stratosphere is ignored entirely. The only nonlinear terms are vertical advection of temperature and humidity. Compared to current GCMs this model contains only minimal physics, but compared to the simple models used for ENSO studies, it is elaborate. Correspondingly, its absorption of computer time falls between the two extremes, but since it is still one or two orders of magnitude faster than a traditional GCM, it can easily be integrated for many years during a normal workday on a commonly available workstation.

Given this efficiency and simplicity it is perhaps surprising that the model does as well as it does. We have shown that the standard model produces a simulation of tropical precipitation and low-level winds that, by the standard of the competitors, is quite reasonable. Indeed the rainfall obtained with this model appears to be better (judged qualitatively) than obtained in many current GCMs. Much of the credit for this lies with the Betts–Miller convection scheme. In particular the later version of this scheme (Betts and Miller 1993) leads to a much better rainfall simulation than was obtained with the earlier version. The later scheme accounts for the low-level drying and cooling by convective downdrafts, which acts to regulate convection and limit the regions in which it occurs. However, numerous problems remain with the rainfall, including areal extent that is still too extensive, an overextended SPCZ, and ITCZs that are too weak. These problems are common to most GCMs. However, the version of the ECMWF model that uses the Betts–Miller scheme has no such problem suggesting that, in our model, the problem lies with other processes or model resolution rather than with the convection scheme itself.

The low-level winds are simulated less successfully than the precipitation. Many of these problems can be traced to absence of orography that frequently allows flow across what would be severe mountain ranges such as the Central America Cordillera or the Andes. Other problems can be traced to errors in the model precipitation and hence thermal forcing. But yet other problems, such as the equatorial easterlies not extending far enough west in the Pacific, are more likely related to neglect of nonlinear accelerations.

The model climate was found to be quite insensitive to the choice of PBL parameterization. Replacing the similarity theory scheme with some very simple assumptions left much of the precipitation and low-level wind field unchanged apart from a reduction of rain over land and a weakening of the Pacific ITCZ, both of which count as deteriorations. It was argued that this follows from the assumption of fixed radiative cooling that requires the same balancing surface fluxes whatever the scheme for their computation. This was corroborated by an experiment in which the model radiation was halved. As predicted by theory this substantially reduced the model precipitation.

The important role that radiation plays in the scale organization of tropical convection was illustrated by an integration that used a clear-sky radiative transfer calculation. Deep convection became ubiquitous in the Tropics, and shallow convection was absent. Cloud-top radiative cooling is essential to maintain shallow convection. The shallow convection, by drying the lowest level and moistening above, allows the surface flux of moisture and subsidence to balance. If shallow convection is absent, deep convective drying must balance the surface flux. This implies that the problem of the scale selection of tropical convection is not just a dynamical one. Cloud–radiation interaction is an essential ingredient. Comparing these results with those for different PBL parameterizations shows how the model climate is more effected by assumptions regarding the large-scale sinks of energy rather than by details of the modeled internal energy exchanges. As such incorporation of a realistic radiation scheme is a necessity and a priority.

Equally important for the next phase of model development is inclusion of orography. We have argued, in accordance with many others, that orography is essential to a reliable simulation of the Asian summer monsoon. Its absence in this model leads to a gross underestimate of the monsoon rainfall. It was discovered, by inclusion of a seasonally varying external heat source in this region, that an improved monsoon also has a modest affect on the trade winds over the tropical Pacific, illustrating some degree of coupling between these two sectors. Inclusion of orography is also expected to improve the modeled circulation in the eastern tropical Pacific.

To conclude, we have presented a climate simulation made with a "minimal physics," linear primitive equation model that is encouraging for those who are interested in development and use of models of intermediate complexity for long-term climate studies. Given its simplicity and exclusion, or oversimplification, of several processes known to be important, it is surprising that it is capable of producing a simulation of certain aspects of tropical climate (such as rainfall) that are not noticeably worse than those of much more complete models. Inclusion of realistic radiation and orography are indicated to be top of the list of priorities for future work.

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