The hydrological Cycle of the Intraseasonal Oscillation

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1. INTRODUCTION

Since better understanding of 30-60 day oscillation can lead to improvements in medium and long range weather prediction, this phenomenon has attracted considerable attention. The 30-60 day oscillation was identified by Madden and Julian (1971, 1972) as the eastward propagation of a wavenumber 1 disturbance in tropical, tropospheric zonal wind and surface pressure fields. The range of the period of this disturbance corresponds to a zonal phase speed of 10 to 15 ms^{-1}. Other observational studies have documented disturbances with a similar period of 30 to 60 days in other physical fields. These findings include the oscillations in atmospheric angular momentum (Anderson and Rosen, 1983, and Risbey and Stone, 1988), convective activities and cloudiness (Yasunari, 1980), outgoing long wave radiation over the Indian-Western Pacific Oceans (Murakami *et al.*, 1986, among others) and perturbations in precipitation amount over the same area (Hartmann and Gross, 1988). In addition to atmospheric phenomena, 30-60 day fluctuations of sea level height (Enfield, 1987, and Mitchum and Lukas, 1987) and sea surface temperature (SST) (Enfield, 1987) have also been found. From these studies it is obvious that the 30-60 day oscillation is a complex process involving disturbances in both the tropical atmosphere and ocean.

Despite the multitude of observational studies, a satisfactory theoretical explanation of the 30-60 day oscillation has yet to be agreed upon. Since Gill's (1982) finding, that the phase speed of the tropical waves can be significantly reduced due to reduction of the static stability (hence the equivalent depth) of the atmosphere by latent heat release, the linear tropical wave theory (Matsuno, 1966) after being refined by the diabatic heating parameterization, has been widely applied in explanation of the 30-60 day oscillation. For example, Lau and Peng (1987) proposed the 'mobile wave CISK' mechanism and Chang and Lim (1988) developed the 'Kelvin wave CISK' theory. Although the selected or the newly generated waves in these theories were indeed amplifying as propagating eastward the phase speed of these waves was crucially dependent on the diabatic heating profile used in these models. In order to have a phase speed close to the observed one the heating profile has to have a peak in the lower troposphere. This heating profile, however, is more representative of situations for the mid-latitude than of the tropics (Reihl and Malkus, 1958).

Attempts were also made to find new modes which could be responsible for the observed 30-60 day oscillation. Anderson and Stevens (1987), for example, specified several slow poleward propagating unstable modes in their 2-D eigenvalue model which included an equatorial symmetric Hadley circulation.

Studies in which the intraseasonal oscillation was regarded as an atmospheric response to a sporadic external heating with a 30-60 day period is another category in the approach. By postulating an oscillatory heating as the forcing, many experiments with different atmospheric models were able to produce the expected low frequency oscillatory features (Yamagada and Hayashi, 1984, and Salby and Gacia, 1987). The lack of an explicit instability mechanism in this category of explanation remains questionable, however, if one views the 30-60 day oscillation as purely an intrinsic atmospheric oscillation.

The sporadic 'external' forcing may turn into an 'internal' one when the ocean and atmosphere are considered as one system. Krishnamurti *et al.* (1988) has found the observational evidence of the air-sea interaction on the time scale of 30-60 days. Emanuel (1987) and Neeling *et al.* (1987) looked at the evaporation-wind feedback effect in an ocean-atmosphere system. They showed that the low frequency oscillation can be amplified by drawing energy from the warm ocean surface through this feedback. Although the existence of the oscillation does not depend on the evaporation-wind feedback (Neeling *et al.*, 1987) their studies certainly provided a different perspective for understanding this problem.

In this study, we propose a new theory explaining the origin of low frequency oscillations in the tropics. In our models both the lower atmosphere and upper ocean layers are included and are treated as a fully coupled system. Based on the model results, we suggest that the observed 30-60 day oscillations in the tropical troposphere and in the ocean are the manifestation of the cloud-radiation, atmosphere-ocean feedback processes, and that their realization depends on the hydrologic cycle in this system.

In the next section, the basic physical mechanism for the 30-60 day oscillation in terms of the hydrologic cycle is presented. Sections 3, 4 and 5 describe three different models and their results. Our conclusions are summarized in Section 6.

2. MECHANISM

The tropical atmosphere above the warm ocean surface is heated through the energy exchanges with the ocean. Meanwhile evaporation from the ocean moistens the lower part of the tropical troposphere. The warmer the ocean mixed layer, the more heat and more moisture are added into this part of the atmosphere.

The removal of the accumulated 'extra' heat and moisture from the tropical atmosphere occurs through deep cumulus convection (Riehl and Malkus, 1958). This convection is crucial in the hydrologic cycle of the tropical atmosphere-ocean system. When explaining the conditional instability in the tropical atmosphere, Ooyama (1969) showed that the vertical temperature and moisture structures in the tropics possess a conditionally unstable character before the onset of convection. However, this instability is realized only when the lower troposphere is sufficiently moistened, through surface evaporation and/or large-scale convergence. This characteristic of the tropical atmosphere is important as it highlights the key role played by the moisture in the occurrence of tropical convection. In our conceptual models, presented in the next two sections, we will set a critical moisture value as the 'model atmospheric moisture tolerance' to represent this thermodynamic feature of the tropical atmosphere. This treatment principally follows the philosophy in parameterizing cumulus convection (Kuo, 1975, and Stevens and Lindzen, 1978) and is based on empirical evidence such as the observed moisture changes in synoptic scales wave disturbances (Reed and Recker, 1971).

When the conditional instability is being released, the convection not only diminishes the extra moisture in the atmosphere through precipitation but, through cloud-radiation interaction, also reduces the sea surface temperature by attenuating the solar radiation energy available at the ocean surface. The result of this is to return the extra water substance in the atmosphere back to the ocean and to reduce the sea surface temperature to its 'climatological' value.

In short, a natural cycle represented by the hydrologic process in the atmosphere-ocean system is spontaneously carried on in the tropics. Our model shows that, simply restricted by the 'moisture tolerance', such a cycle possesses an intraseasonal time scale.

3. A SIMPLE MODEL WITH FIXED SST

a) The Model

In this section we demonstrate the existence of the hydrologic cycle in a simple atmosphere-ocean system with *fixed* ocean parameters. The schematic structure of this model is given in Fig. 1. As shown in Fig. 1 only thermodynamic process are considered in this model. Our attention is to describe the accumulation of the extra water content in the atmospheric mixed layer through evaporation and the removal of this content by precipitation; horizontal convergence and divergence are neglected.

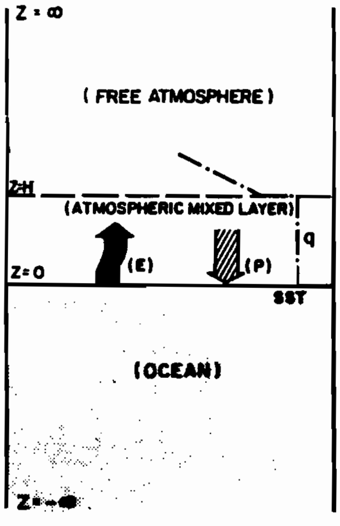

Fig. 1: Schematic structure of the model (MD1) with a fixed sea surface temperature (SST).

The temperature of the atmospheric mixed layer is characterized by a near-surface atmospheric temperature $T_a = T_s - \Delta T(°C)$. Precipitation is assumed to take place when the 'atmospheric moisture tolerance' is achieved. As an initial approach we specify this critical condition as that when the mixing ratio of the atmospheric mixed layer reaches 90% of the saturation value at its temperature, T_a.

The amount of moisture evaporated in each time interval, Δt, is calculated by

$$\Delta E = \int_{\Delta t} E(t)dt\,, \tag{1}$$

where,

$$E(t) = \rho_a C_D |u_s| |q_{sat}(T_s) - q_{atm}(t)| \tag{2}$$

(Neeling *et al.* 1987). In (2), ρ_a is the mean density of the atmospheric mixed layer, C_D is the surface drag coefficient and is treated as a constant of 0.001. The parameter u_s is the low level wind at 10 meters above the sea surface, $q_{sat}(T_s)$ is the saturation mixing ratio of the atmosphere at temperature T_s and $q_{atm}(t)$ is the actual mixing ratio of atmospheric mixed layer at time t. The value of q_{atm} is updated at time step $t + \Delta t$ by an increment Δq given by

$$\Delta q = \frac{1}{(\rho_a H)} \int_{\Delta t} (E(t) - P)dt\,. \tag{3}$$

Here P represents the precipitation rate and is assumed as a constant, $2.5mm/day$ (Wang, 1988) during the model's rainy period and H is the mixed layer depth.

By using a mean atmospheric moist state and the precipitation rate, the model is integrated with a one hour time interval for 1000 days.

b) Model Result and Sensitivity

Fig. 2 shows the results of two model integrations. Model parameters used in these integrations are listed in the figure caption. The displayed field, extra water content, indicates the column moisture amount above an assumed background state, which corresponds to a mixing ratio (q_{atm}) of $15g/kg$ in the atmospheric mixed layer.

An *ad hoc* parameterization of the low-level wind, intended to represent the higher winds and associated transfers between atmosphere and ocean during convection,

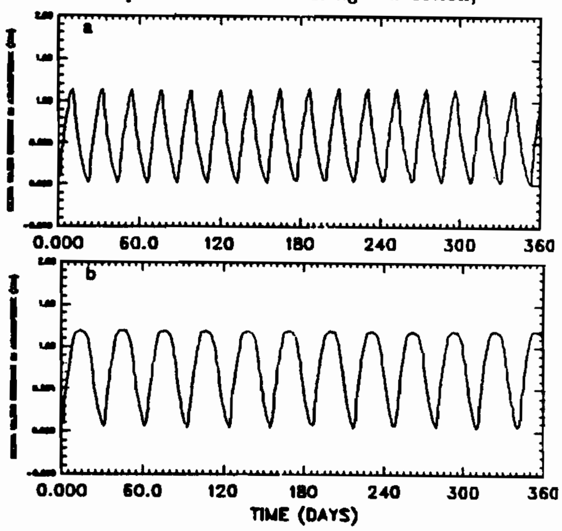

Fig. 2: Predicted hydrologic circulation in the model atmosphere-ocean system. The model low level wind, u_s, used in (a) is a constant of 2.0 m/s, and in (b) is that given in Eq. (4), with $u_b = 2\ m/s, u_m = 12\ m/s$, and $\Delta t_d = 13 days$. Other model parameter values are $T_s = 28.5°C$, $T_a = 28.0°C$, $H_{atm} = 1500m$ and $P = 2.5mm/day$.

$$u_s(t) = \begin{cases} u_b & t \le t_c \\ u_m & t_c \le t \le t_c + \Delta t_m \\ u_m - \frac{u_m - u_b}{\Delta t_d}[t - (t_c + \Delta t_m)], & t_c + \Delta t_m \le t \le t_c + \Delta t_m + \Delta t_d, \end{cases} \quad (4)$$

is used in getting the result in Fig. 2b. In this expression u_b is the mean background wind speed and t_c is the time when the atmospheric moisture tolerance is achieved. The maximum wind speed, u_m is assumed to occur during the prescribed time interval Δt_m when deep cumulus convection is present. This time increment is prescribed to be 2 days,

From the results shown in Fig. 2 the existence of the hydrologic cycle operating in this simple atmosphere-ocean system is clearly evident. Through the processes of evaporation and precipitation, which are constrained by the moisture tolerance and its implied conditional instability, the hydrologic cycle is naturally introduced in this system. Furthermore, through a comparison of these two results, we see that the presence of this cycle is independent of the detailed structure of the low level wind. This independence suggests that the oscillation in the low level wind field may be considered a consequence of the hydrologic cycle. Since the time evolution of the heating profile in the tropical troposphere is directly related to the different phases of the hydrologic cycle, this cycle provides an external heat/energy source for the atmosphere. This source disturbs the energy state of the tropical atmosphere to which the wind and mass fields of the atmosphere then adjust. Thus this hydrologic cycle is a plausible mechanism for initiating oscillations of similar frequency in wind and mass fields in the tropical atmosphere. The influence of the varying low level wind field on the frequency of the hydrologic cycle is also noticeable in Fig. 2, although it is does not determine the existence of the cycle *per se*.

Results of model integrations with different model parameters are presented in Fig. 3. These results were obtained with the same low level winds as used to produce Fig. 2b. A significant characteristic shown in this figure, is that the period of the resultant hydrologic cycle in this system increases with increasing model SST. This result is consistent with the observation that low frequency oscillations are preferred in regions of warmest SSTs (Webster, 1987).

The model sensitivities to different low level wind speeds and values of precipitation rate are also illustrated in Fig 3.

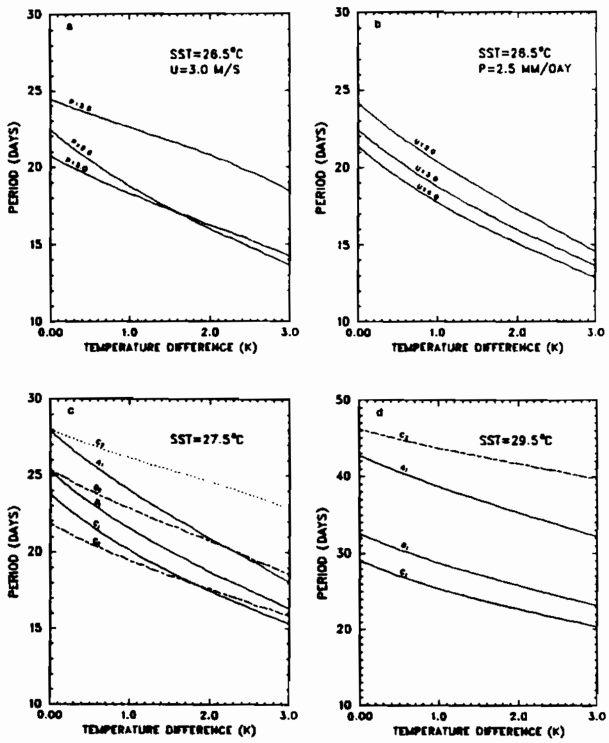

Fig. 3: Predicted period of hydrologic circulation with different model conditions. The abscissa shows the temperature difference between the ocean and the atmosphere. (a) and (b) show the model sensitivities to the changes in precipitation rate P [unit: mm/day] and the model low level wind u_s [unit: m/s], respectively. In (c) and (d), the 3 groups of results, e.g. A_i, B_i and C_i are for 3 different P's used in the model, i.e. 2.5, 3.0 and 3.5 mm/day. The subscript i, from 1 to 3, stands for different low level wind conditions of 2.0, 3.0 and 4.0 m/s. Lines are not plotted in (c) and (d) when a balance between E and P is the case and therefore no oscillation occurs.

4. COUPLED INTERACTIVE ATMOSPHERE-OCEAN MIXED LAYER MODEL

The existence of the hydrologic cycle and its relationship to the low frequency oscillation in the tropical troposphere have been shown by the one-way interaction model (hereafter MD1) of Section 3. Since a fixed SST was applied in MD1, the question of how this cycle as well as its time scale is influenced by a time varying SST still remains. In this section we address some aspects of this issue.

a) The Model

The model structure and the major physical processes are schematically illustrated in Fig. 4. As in MD1, we focus on the thermodynamics of this atmosphere-ocean system and ignore any detailed dynamics. The model's ocean mixed layer process follows that in Kraus and Turner (1967), and Denman (1973)

In this model to incorporate in the model SSTs which vary in time, a detailed energy budget in the ocean mixed layer is required. As the only energy source in this budget, the input solar energy at the ocean surface needs to be included. This input further varies with presence or absence of cumulus convections through cloud-solar radiation interaction. On the other hand, the energy losses through infrared radiation (IR) from ocean surface, and sensible and latent heat exchanges with the atmosphere are also considered. Since explicit prediction of key processes, such as cloud generation and interaction of clouds with solar radiation, is beyond the scope of this mechanistic study, we parameterize their effects. To express the cloud influence on the solar radiation, for example, we assume different values of the solar irradiance at the surface under varying convective cloud cover conditions. These values are obtained from observations and theoretical studies. A similar method is also applied to represent the IR cooling process. The main reason for neglecting the detailed treatment of these processes is to avoid the uncertainties associated with complex processes involving multiple scales, some of which are poorly understood, and therefore to concentrate more on the basic physical mechanisms.

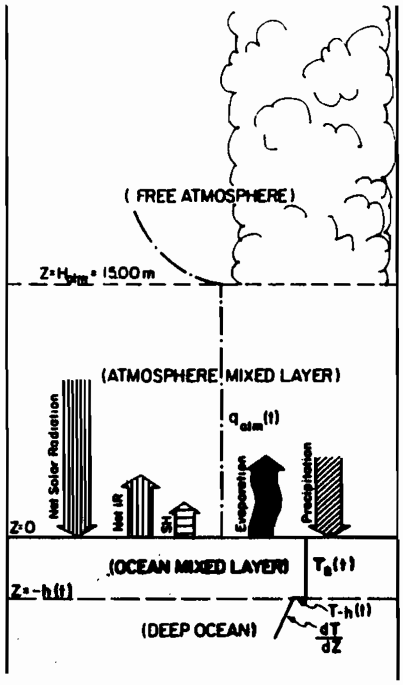

Fig. 4: Schematic structure of the fully coupled atmosphere-ocean mixed layer model (MD2).

In the model atmosphere, the process is similar to MD1. The moisture amount transferred into the model atmospheric mixed layer is calculated again using the bulk aerodynamic formula

$$E(t) = \rho_a C_D |u_*(t)| \left(q_{sat}[T_s(t)] - q_{atm}(t)\right). \qquad (5)$$

Here, $q_{sat}[T_s(t)]$ is time dependent as SST varies with time in this model. A Bowen ratio of 0.2 is used to obtain the value of the sensible heat flux required as an input to the ocean model.

The solar irradiance at the ocean surface are specified by the step function

$$R_{sw} = \begin{cases} R^n_{sw}, & \text{non-convective;} \\ R^c_{sw}, & \text{convective.} \end{cases} \qquad (6)$$

The values in (6) are based upon the satellite and observational data analyses by Gautier and Katsaros (1984), Gautier (1986) and Gautier (1988). With these values, the bulk features of the influence of deep cumulus convection on the solar radiation transfer in the atmosphere have been included. Likewise the step function for IR change is

$$R_{IR} = \begin{cases} R^n_{IR}, & \text{non-convective;} \\ R^c_{IR}, & \text{convective,} \end{cases} \qquad (7)$$

where the values for R^n_{IR} and R^c_{IR} are based on a study by Stephens *et al.* (1981).

b) Model Results

Table 1 lists some of the model experiments we conducted. The physical parameters, their values, and the model predicted quantities are included in this table.

Since only thermodynamic and turbulent mixing processes in the two mixed layers are considered, only a few of these experiments reach a steady or an equilibrium state through the period of integration. Fig. 5 shows the result from one of the experiments (experiment 1 in Table 1) in which an equilibrium state has been reached. From this figure we see that the oscillation of the extra water content in the atmospheric mixed layer still experienced when the model has varying SST. This oscillation, through the coupling and feedback with the solar radiation and hence the surface energy process, influences the processes in the ocean mixed layer (Figs. 5b,c and d). The time scale of these oscillations predicted by the model is 33 days.

As we have discussed in Section 2, in the process of the hydrologic circulation a pulsating energy disturbance is released in the convection phase. This disturbance then causes dynamic perturbations in the atmospheric momentum and mass fields (Salby and Garcia, 1987). The observed oscillation in the atmospheric relative angular momentum (Anderson and Rosen, 1983) may also be related with this pulsating energy disturbance. On the other hand, this energy disturbance, in addition to the change in the solar energy input, affects the energy budget of the ocean mixed layer. As a consequence of the hydrologic circulation in the atmosphere-ocean system, the 30-60 day oscillations therefore appear in both the tropical atmosphere and ocean.

Because the thermodynamic processes depend on SST in a highly nonlinear fashion, the results illustrated in Table 1 differ from those obtained from MD1. In Fig. 3, a higher SST in the one-way model causes a dramatic increase in the period of the model's hydrologic circulation. Results from the fully coupled model, however, are not as simple. Using $29°C$ as the initial SST, the hydrologic cycle reaches a steady period of 35 days (experiment 2 in Table 1). This period is quite close to that predicted in experiment 1 where the initial SST is $28°C$. However when a colder SST is used to start the model run, the predicted result are distinctly different (Fig. 6).

Results presented in Table 1 also show the model sensitivity to other model parameters.

5. ONE-DIMENSIONAL GCM AND RESULTS

To test our hypothesis in a more complete manner different experiments have also been conducted with the one-dimensional version of the UCLA/CSU GCM (Randall and Dazlich, 1989). This 1-D model includes the full radiation and moist physics parameterization of the GCM, and is coupled to a slab of ocean of fixed depth. The slab ocean does not exchange energy with deeper ocean layers; its temperature is controlled entirely by the surface energy flux.

At this stage some very preliminary results from experiments with this 1-D GCM model have been obtained and are shown in Figs. 7 and 8. These results support the major findings from our previous conceptual models. When run with a ocean slab depth of $60m$, the model produces obvious oscillation of sea surface temperature, with amplitude $0.4°C$ and a period of 60 days (Figs. 7a and 8a). This oscillation is accompanied by 60-day fluctuations of the surface energy flux, precipitation rate and cloudiness (Figs. 7b, 7c and 8b). The changes in the surface energy flux are due to changes in the absorbed solar radiation, which are controlled by the cloudiness fluctuations or model convections.

During the warming phase of the oscillation, the sea surface temperature and the convective precipitation rate gradually increase. Eventually the cloudiness increases to the point that the solar radiation absorbed by the ocean is reduced, allowing the ocean to begin cooling. This cooling is followed by a reduction in the model convective activity, which leads to a reduction in the cloudiness. The absorbed solar radiation then increases, and the model hydrologic cycle is re-initiated.

Table 1: Model Experiments and Results – Among the output quantities, τ is the predicted period of the model hydrological circulation. Σq is the maximum extra water amount accumulated in the model atmosphere before convection starts. ΔT and Δh are respectively the amplitudes of changes in the SST and depth of ocean mixed layer in a hydrological circulation. When a steady state is not reached in a model run, these quantities are given from a nominal cycle in model integration.

Physical Parameter	Experimental Value					
	1	2	3	4	5	6
R^n_{SW} (W/m^2)	250	250	250	250	250	260*
R^c_{SW} (W/m^2)	120	120	120	120	120	105*
R^n_{IR} (W/m^2)	40	40	40	40	40	40
R^c_{IR} (W/m^2)	40	40	40	40	40	40
u_s : $(u_b, u_m, \Delta t_d)$	(2, 12, 13.0)	(2, 12, 13.0)	(2, 12, 13.0)	(2, 12,11.5*)	(2, 12, 15.0*)	(2, 12, 13.0)
$P\,(mm/day)$	2.5	3.0*	2.5	2.5	2.5	2.5
$T_s(t=0)(°C)$	28.0	29.0*	26.0*	28.0	28.0	28.0
$T_{-h}(t=0)(°C)$	25.0	26.0*	23.0*	25.0	25.0	25.0
Output Quantity	Model Predicted Result					
τ (days)	33	35	the 5th: 23 the 9th: 15	the 5th: 33 the 20th: 37	the 5th: 33 the 20th: 26	31
$\Sigma q\,(cm/m^2)$	1.49	1.51	the 9th: 0.35	the 20th: 1.70	the 20th: 0.80	1.40
$\Delta T_s(°C)$	1.68	1.63	the 9th: 0.83	the 20th: 1.67	the 20th: 1.41	1.83
$\Delta h(m)$	21.85	21.39	the 9th: 13.51	the 20th: 21.48	the 20th: 19.50	24.93

*: Parameter value different from that in the controlled experiment (experiment 1).

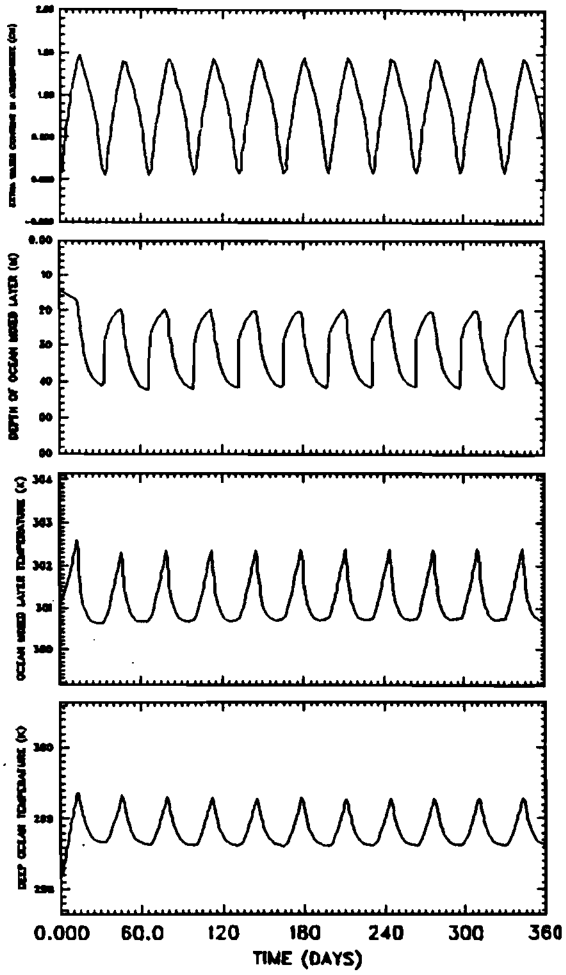

Fig. 5: Model results from experiment 1 in Table 1.

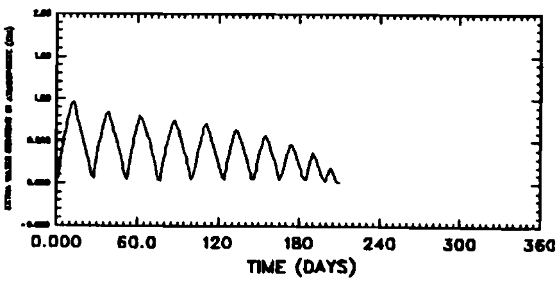

Fig. 6: Temporal variation of the extra water content in the model atmosphere from experiment 3 in Table 1.

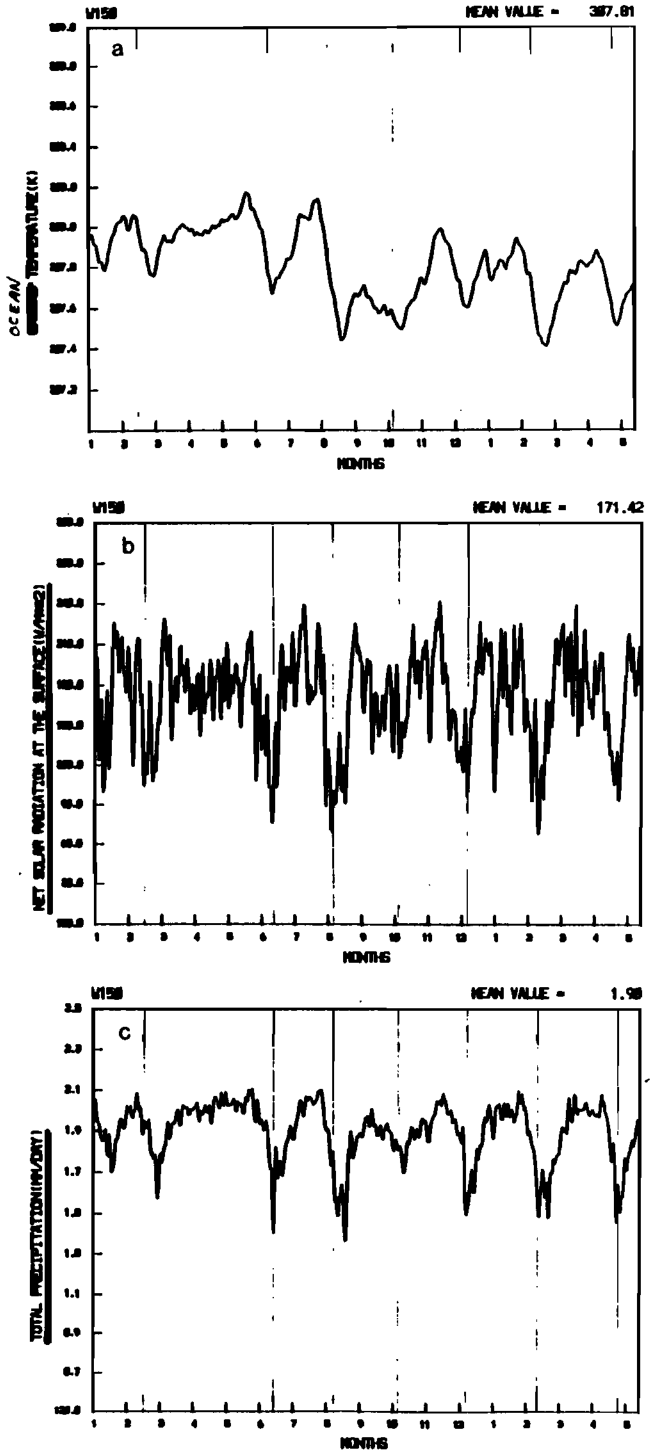

Fig. 7: Temporal variations of (a) SST, (b) solar radiation absorbed at the ocean surface, and (c) daily precipitation from the 1-D GCM.

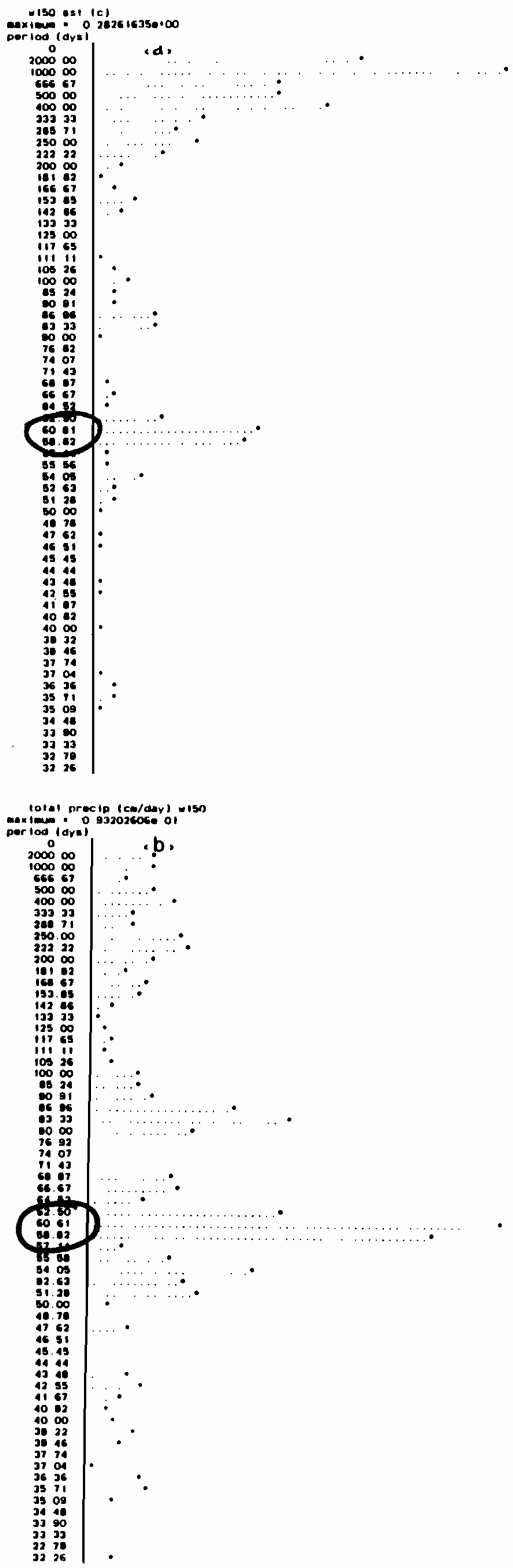

Fig. 8: Square of the amplitudes of the Fourier harmonics of (a) model SST, and (b) model precipitation. Peaks at about 60 days are outstanding.

6. CONCLUDING REMARKS

In this study we have shown that the 30-60 day oscillations in the tropical atmosphere and ocean can result from the hydrologic circulation in a coupled atmosphere-ocean system. This circulation is intrinsic to this system and is a consequence of the energy exchange and energy balance in this system.

The time scale of this circulation can be determined in a simple coupled atmosphere-ocean model when the role of moisture played in the atmosphere is incorporated in the model. By parameterizing this role in terms of the 'model atmospheric moisture tolerance' in our conceptual models we showed that the hydrologic cycle is spontaneously carried on in the tropical atmosphere-ocean system. Its existence was basically independent of the variations of the low level wind.

Our hypothesized relationship between the hydrologic circulation and the intraseasonal oscillation in the tropical atmosphere and ocean was also examined in a 1-D GCM. The model results were positive in supporting this relationship. Meanwhile, the analysis of the thermodynamic process in the 1-D GCM illustrated the mechanism we proposed.

Based on our results we conclude that the hydrologic cycle in the tropical atmosphere-ocean system is fundamental to the observed intraseasonal oscillations in the tropics. The physical restriction of the accumulation of latent heat in the lower tropical troposphere and the reduction of the source supply of energy to the tropical atmosphere-ocean system, through cloud-solar radiation interaction, are crucially important to the presence and time scale of this hydrologic cycle.

7. ACKNOWLEDGMENTS

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8. REFERENCES

Anderson, J.R., and R.D. Rosen, 1983: The latitude–height structure of 40– 50 day variations in atmospheric angular momentum. *J. Atmos. Sci.*, **40**, 1584–1591.

Anderson, J.R. and D.E. Stevens, 1987: The presence of linear wavelike modes in a zonally symmetric model of the tropical atmosphere. *J. Atmos. Sci.*, **44**, 2115–2127.

Anderson, J.R. and D.E. Stevens, 1987: The response of the tropical atmosphere to low-frequency thermal forcing. *J. Atmos. Sci.*, **44**, 676–686.

Chang, C.-P., and H. Lim, 1988: Kelvin Wave-CISK: A possible mechanism for the 30–50 day oscillations. *J. Atmos. Sci.*, **45**, 1709–1720.

Denman, K.L., 1973: A time–dependent model of the upper ocean. *J. Phys. Oceanogr.*, **3**, 173–184.

Emanuel, K.A., 1987: An air–sea interaction model of intraseasonal oscillations in the tropics. *J. Atmos. Sci.*, **44**, 2324–2340.

Enfield, D.B., 1987: The intraseasonal oscillation in eastern Pacific sea levels: How is it forced? *J. Phys. Oceanogr.*, **17** 1860-1876.

Gautier, C., 1986: Evolution of the net surface short wave radiation over the Indian ocean during summer MONEX (1979): A satellite description. *Mon. Wea. Rev.*, **114**, 525-533.

Gautier, C. and K. Katsaros, 1984: Insoluation during STREX. Part I: Comparisons between surface measurements and satellite estimates. *J. Geophys. Res.*, **89**, 11779-11788.

Gautier, C., 1988: Surface solar radiance in the central Pacific during Tropic Heat: Comparisons between *in situ* measurements and satellite estimates. *J. Climate*, **1**, 600-608.

Gill, A.E., 1982: Studies of moisture effects in simple atmospheric models: the stable case. *Geophys. Astrophys. Fluid Dynamics*, **19**, 119-152.

Hartmann, D.L., and J.R. Gross, 1988: Seasonal variability of the 40-50 day oscillation in wind and rainfall in the tropics. *J. Atmos. Sci.*, **45**, 2680-2702.

Kraus, E.B., and J.S. Turner, 1967: A one-dimensional model of the seasonal thermocline II: The general theory and its consequences. *Tellus*, **19**, 98-106.

Krishnamurti, T.N., D.K. Oosterhof, and A.V. Mehta, 1988: Air-sea interaction on the time scale of 30-50 days. *J. Atmos. Sci.*, **45**, 1304-1322.

Kuo, H.-L., 1975: Instability theory of large-scale disturbances in the tropics. *J. Atmos. Sci.*, **32**, 2229-2245.

Lau, K.M., and L. Peng, 1987: Origin of low-frequency (intraseasonal) oscillations in the tropical atmosphere. Part I: Basic theory. *J. Atmos. Sci.*, **44**, 950-972.

Madden, R., and P.R. Julian, 1971: Detection of a 40-50 day oscillation in the zonal wind in the tropical Pacific. *J. Atmos. Sci.*, **28**, 702-708.

Madden, R., and P.R. Julian, 1972: Description of global-scale circulation cells in the tropics with a 40-50 day period. *J. Atmos. Sci.*, **29**, 1109-1123.

Matsuno, T., 1966: Quasi-geostrophic motions in the equatorial area. *J. Meteor. Soc. Japan*, **44**, 25-42.

Mitchum, G.T., and R. Lukas, 1987: The latitude-frequency structure of Pacific sea level variance. *J. Phys. Oceanogr.*, **17**, 2362-2365.

Murakami, T., L. Chen and A. Xie, 1986: Relationship among seasonal cycles, low-frequency oscillations, and transient disturbances as revealed from outgoing longwave radiation data. *Mon. Wea. Rev.*, **114**, 1456-1465.

Neeling, J.D., I.M. Held and K.H. Cook, 1987: Evaporation-wind feedback and low-frequency variability in the tropical atmosphere. *J. Atmos. Sci.*, **44**, 2341- 2348.

Ooyama, K., 1969: Numerical simulation of the life cycle of tropical cyclones. *J. Atmos. Sci.*, **26**, 3-40.

Randall, D. A. and D. Dazlich, 1989: The *CSU* general circulation model: *A user's guide.* informal manuscript.

Reed, R.J. and E.E. Recker, 1971: Structure and properties of synoptic-scale wave disturbances in the equatorial western Pacific. *J. Atmos. Sci.*, **28**, 1117-1133.

Riehl, H., and J.S. Malkus, 1958: On the heat balance in the equatorial trough zone. *Geophysica*, **6**, 3-4.

Risby, J.S., and P.H. Stone, 1988: Observations of the 30-60 day oscillation in zonal mean atmospheric angular momentum and high cloud cover. *J. Atmos. Sci.*, **45**, 2026-2038.

Salby, M.L., and R.R. Garcia, 1987: Variations induced by interference of stationary and traveling planetary waves. *J. Atmos. Sci.*, **44**, 2679-2711.

Stephens, G.L., G.G. Campbell and T.H. Vonder Haar, 1981: Earth radiation budgets. *J. Geophys. Res.*, **86**, C10, 9739-9760.

Stevens, D.E. and R.S. Lindzen, 1978: Tropical wave CISK with a moisture budget and cumulus friction. *J. Atmos. Sci.*, **35**, 940-961.

Wang, B., 1988: Dynamics of tropical low-frequency waves: An analysis of the moist Kelvin wave. *J. Atmos. Sci.*, **45**, 2051-2065.

Webster, P.J., 1987: The role of hydrological processes in ocean-atmosphere interaction. USTOGA 8, 71-100.

Yamagada, T., and and Y. Hayashi, 1984: A simple diagnostic model for the 30-50 day oscillation in the tropics. *J. Meteor. Soc. Japan*, **62**, 709-717.

Yasunari, T., 1980; A quasi-stationary appearance of 30-50 day period in the cloudiness fluctuations during the summer monsoon over India. *J. Meteor. Soc. Japan*, **58**, 225-229.

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Nouméa, New Caledonia
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PROCEEDINGS

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