A simple thermodynamic model for seasonal variation of monsoon rainfall

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Monsoons occur in the tropics on account of complex interaction between the continents, the ocean and the atmosphere and hence no simple model of the monsoon has been proposed so far. In this paper, a simple diagnostic model of the monsoon has been proposed based on the constraints imposed by energy and the moisture balance in a vertical column of the tropical atmosphere. The model demonstrates that the monthly mean rainfall depends upon evaporation, the net energy available in the atmosphere and the vertical stability of the atmosphere. The model simulates the seasonal variation of rainfall in India, Africa and South America remarkably well without explicitly invoking the concept of land–sea contrast in temperature. The model is useful to understand the impact of deforestation, greenhouse gases and aerosols on monsoon rainfall.

THE understanding and prediction of seasonal variation of the rainfall in the tropics has been a major goal in tropical meteorology. During the past twenty-five years, complex General Circulation Models (GCMs) of the atmosphere have been used to simulate the seasonal variation of precipitation in the tropics[1]. Although these models exploited the ability of the computer to perform millions of computations per second, they have not been very successful in simulating accurately the seasonal variation of precipitation during the monsoon[2]. This could be on account of poor representation of processes such as rainfall, cloud–radiation interaction, evaporation from vegetation and flow around mountains. On account of the inherent complexity of a GCM, it is difficult to identify the factors responsible for the poor simulation of monsoon rainfall by a GCM. Ironically, these complex models constructed by human beings are as inscrutible as nature. Hence a simple diagnostic model of the monsoon is necessary to identify the factors that govern the seasonal variation of rainfall. We discuss in this paper a simple diagnostic model based on the energy and moisture balance of the tropical atmosphere.

In the tropics, horizontal gradients of temperature and moisture are weak and hence their contribution to the energy balance can be neglected when compared to the contribution of the terms such as radiation, latent heat release or vertical gradient of temperature. In the tropics,

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the integral form of the law of conservation of energy and moisture can be written as (for details, see Neelin and Held[3])

$$\int_0^1 \omega\, [\partial m/\partial p^*]\partial p^* = g[F_B - F_T], \tag{1}$$

$$\int_0^1 \omega\, [\partial Lq/\partial p^*]\partial p^* = gL[E - P], \tag{2}$$

where ω is the vertical velocity; m is moist static energy $= s + Lq$; s is dry static energy; q is specific humidity; p is pressure; p_o is surface pressure; p^* is p/p_o; g is acceleration due to gravity; L is latent heat of condensation; E is evaporation from the soil; F_B is sum of radiative, sensible and evaporative heat fluxes at the surface; F_T is radiative flux at the top of the atmosphere; and P is precipitation.

Equations (1) and (2) can be combined to obtain the following simple expression for precipitation:

$$L[P - E] = \{F_B - F_T\}/\{\delta - 1\}, \tag{3}$$

where

$$\delta = -\{\int_0^1 \omega\, [\partial s/\partial p^*]\partial p^*\} / \{\int_0^1 \omega\, L[\partial q/\partial p^*]\partial p^*\}.$$

On monthly mean scales the energy stored in the soil in the continents is small and hence net flux at the bottom of the atmosphere (i.e. F_B) is close to zero[4]. The net flux at the top of the atmosphere (F_T) is purely radiative. This flux is measured by satellites and is known as net radiation at the top of the atmosphere (Q_{net}). By definition, $Q_{net} = -F_T$.

In tropical continents, the above equation can, therefore, be simplified to

$$P = E + Q_{net}/\{\delta - 1\}. \tag{4}$$

In the above equation P, E, Q_{net} can be expressed in terms of mm/day or W/m^2. The second term on the right-hand side of eq. (4) represents the contribution to rainfall on account of convergence of moisture. Most meteorologists estimate this term by evaluating the divergence of the horizontal wind. We have estimated the convergence of moisture on the basis of the constraints imposed by energy available in a vertical column of the tropical atmosphere. In eq. (4), the parameter δ is a measure of the vertical stability of the atmosphere. During the dry season δ is large, while during the wet season, δ is small. The parameter δ is, however, always greater than 1 and hence the sign of the second term in eq. (4) is determined by the sign of Q_{net}. In regions of the tropics wherein Q_{net} is negative (i.e. winter) the amount of rainfall is less than evaporation. In regions wherein Q_{net} is positive (i.e. summer), the magnitude of the second term is determined by how close δ is to 1. We estimate the value of δ by making simple assumptions regarding the nature of variation of vertical velocity, temperature and specific humidity with height. In regions with heavy rainfall there is a strong ascent and the vertical velocity attains its maximum around the middle of the troposphere, while in the regions of descent (where there is no rainfall) the maximum velocity occurs much closer to the surface. We assume that the maximum in the vertical velocity occurs at $p^* = 0.5$ in all regions. This assumption may not be strictly valid in all regions of the tropics, but we will show later that this assumption does not lead to large errors.

The vertical velocity is assumed to vary with pressure as follows:

$$\omega(p^*) = 4\omega_m p^* [1 - p^*], \tag{5}$$

where ω_m is the value of ω at $p^* = 0.5$.

The variation of specific humidity with pressure is assumed to be

$$q(p^*) = q_o\{p^*\}^{\lambda},$$

where λ is the non-dimensional scale height of water vapour.

If we assume that the tropical atmosphere has a constant temperature lapse rate (Γ), the dry static energy (s) can be written as

$$s = C_P T_o\{p^{*\beta} + a[1 - (p^*)^{\beta}]\},$$

and the moist static energy can be written as

$$m = C_P T_o\{p^{*\beta} + a[1 - p^{*\beta}]\} + Lq_o(p^*)^{\lambda},$$

where a is $g/C_P\Gamma$; β is $R\Gamma/g$; R is ideal gas constant of air; C_P is specific heat of air; T_o is surface temperature, and q_o is surface specific humidity.

With the above assumptions, we obtain the following simple expression for δ.

$$\delta = [p_o(a - 1)\beta(\lambda + 2)C_p T_o]/[Lg p_w \lambda(\beta + 1)(\beta + 2)]. \tag{6}$$

In the above equation, the following expression for the vertically integrated water vapour in the atmosphere (p_w) has been used.

$$p_w = [p_o q_o]/[g(\lambda + 1)].$$

From eq. (6) we find that δ depends upon parameters such as integrated water vapour, surface pressure, surface temperature, temperature lapse rate and the water vapour scale height. Among these, the most important parameter is the integrated water vapour, p_w, while the other parameters do not alter the value of δ substantially. This can be shown by evaluating δ for a range of values of p_o, T_o, λ, β and p_w that is usually encountered in the tropics. We have evaluated δ from the National Centre for Environmental Prediction (NCEP) reanalyses data[5] (which can be considered as a proxy for observations) in July 1988. The variation of δ is shown as function of p_w in Figure 1. We find that most of the variation δ with p_w is captured well by the expression $\delta = C/p_w$, where C is a constant with a value around 85. If this expression is used in eq. (4), we obtain the following simple equation for precipitation in the tropics.

$$P = E + Q_{net}/\{[C/p_w] - 1\}. \quad (7)$$

In the eq. (7), P, E, Q_{net} can be expressed in terms of mm/day or W/m^2. Equation (7) indicates that the continental precipitation in the tropics is controlled primarily by three parameters, namely surface evaporation, net radiation at the top of the atmosphere and the integrated water vapour.

The monthly mean precipitation can now be estimated using eq. (7), if the three terms on the right-hand side of that equation can be measured or estimated accurately. The net radiation at the top of the atmosphere was measured accurately by the Earth Radiation Budget Experiment (ERBE)[6]. Data are available on 2.5° by 2.5° grids for the period 1985–1989. The other two quantities (i.e. E and P_w) can been obtained from the NCEP reanalyses which can be considered as a reasonably good proxy for observations[5]. We can then estimate the monthly mean precipitation in 2.5° by 2.5° grids using the data from ERBE and NCEP reanalyses. In Figure 2, the spatial variation of precipitation over the monsoon regions in the tropics obtained from eq. (7) has been compared with observations obtained from Xie and Arkin[7]. The spatial pattern of rainfall is captured well, but the model does not simulate rainfall accurately in regions with mountains. This could be on account of the simple assumption made with regard to the variation of vertical velocity with pressure. Figure 3 compares the seasonal variation of rainfall in large tropical continental regions simulated by eq. (7) with observations. We find that the simple diagnostic model is able to simulate the seasonal variation of precipitation remarkably well. Note that the simple model estimates accurately the monsoon rainfall in the Indian subcontinent in 1988, but overestimates the monsoon rainfall in 1987. The poor performance of the simple model in 1987 could be on account of limitations of the model or errors in the value of P_w obtained from NCEP reanalysis. A 10% error in the estimate of P_w can result in an error of 20% in the estimate of rainfall. The comparison of P_w obtained from satellite data with NCEP reanalysis indicates that errors in P_w obtained from NCEP reanalysis can be as high as 10%. The accuracy of the simple model is poorer in the dry season than in the wet season. This is understandable because the simple expression we have used for the variation of vertical velocity with pressure (eq. (5)) cannot be expected to be accurate in regions of descent.

The simple model presented here demonstrates that the monthly mean precipitation in the tropical continents depends primarily upon three parameters, namely evaporation, net radiation at the top of the atmosphere and the total water vapour in the atmospheric column. This may

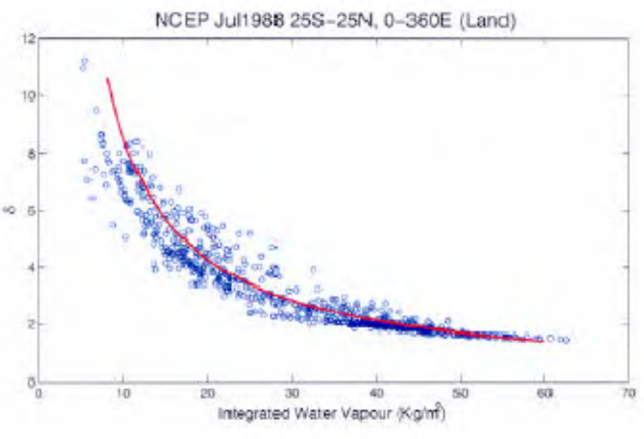

Figure 1. Variation of stability parameter (δ) as a function of the vertically integrated water vapour (p_w) in continental regions in July 1988. δ was calculated from eq. (6). The parameters appearing in the right-hand side of eq. (6) were evaluated from NCEP reanalyses data[5]. Solid line represents the curve $85/p_w$.

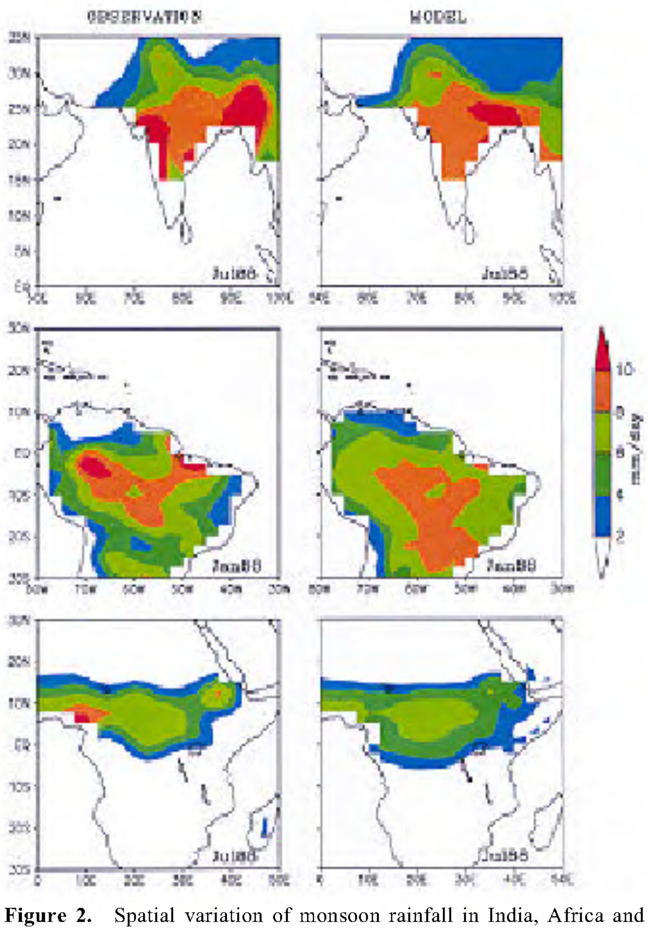

Figure 2. Spatial variation of monsoon rainfall in India, Africa and South America from simple model and observations.

seem to be counter-intuitive to most meteorologists in view of the impact of changes in large-scale circulation during El Nino on the precipitation in the tropics. The large-scale circulation changes can, however, have an impact on local evaporation (through changes in surface wind), local net radiation (through changes in cloudiness) and local integrated water vapour (through changes in vertical velocity). The winds influence precipitation in this model through its impact on local evaporation and local integrated water vapour content of the atmosphere. In continental regions, the integrated water vapour in the atmosphere increases gradually during the pre-monsoon season because winds bring moist air from the ocean to the continent. After the onset of the monsoon, evaporation from the continental surface is determined primarily by the surface winds.

In the recent years there has been a great concern regarding the anthropogenic influence on the radiative processes in the atmosphere. The simple model proposed here indicates clearly the impact of radiative processes on monsoon. The impact of radiation on monsoon rainfall is contained in the term Q_{net} in eq. (7). This term can be written as

$$Q_{\mathrm{net}} = S(1-\alpha) - F^{\uparrow}, \qquad (8)$$

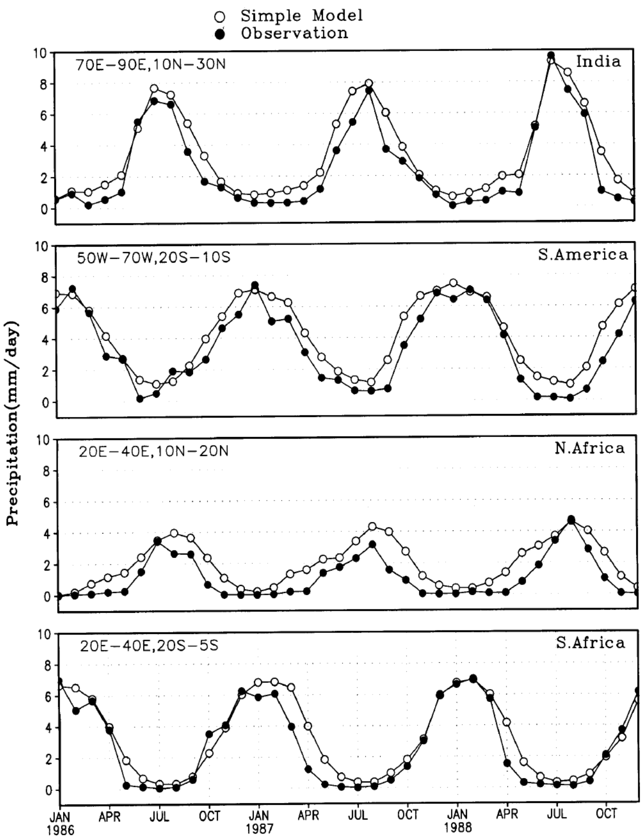

Figure 3. Seasonal variation of monsoon rainfall in tropical continents during 1986 to 1988 from simple model and observations.

where S is solar radiation incident at the top of the atmosphere; α = albedo = reflectivity of the earth–atmosphere system; and $F^{\uparrow}$ is Outgoing Longwave Radiation (OLR).

The impact of increase in CO_2 in the atmosphere on monsoon rainfall can be understood through eqs (7) and (8). On account of increase in CO_2 in the atmosphere, OLR will decrease and this will result in an increase in Q_{net} (through eq. (8)) and hence an increase in rainfall (through eq. (7)). The increase of CO_2 from 300 to 600 ppm will decrease OLR by 4 W/m^2. If there are no other changes (on account of feedback processes), this decrease in OLR will increase net radiation at the top of the atmosphere by 4 W/m^2. The net radiation at the top of the atmosphere is around 80 W/m^2 during the Indian monsoon season. Hence an increase in net radiation by 4 W/m^2 will increase rainfall by about 5%. Most GCMs indicate a much higher increase in monsoon rainfall on account of the doubling of CO_2. This is because of feedback processes that increase the integrated water vapour p_{w} and evaporation. In addition to increase in CO_2 in the atmosphere by human activities, there has been an increase in aerosols. The increase in sulphate aerosols will increase the albedo and hence decrease the net radiation. This will result in a decrease in rainfall. The incoming solar radiation (S) does not vary much on short time scales (i.e. decadal to century), but can vary substantially on longer time scales (thousands to million years) on account of changes in the earth–sun geometry. The simple model proposed here will be useful to understand the impact of changes in S on palaeo-monsoons.

The simple diagnostic model proposed here provides new physical insight regarding the impact of deforestation on rainfall. We can examine the impact of changes in vegetation by perturbing eq. (7) which is then modified to:

$$\Delta P = \Delta E + [P - E]\{[\Delta p_{\mathrm{w}}/p_{\mathrm{w}}][C/(C - p_{\mathrm{w}})] + \Delta Q_{\mathrm{net}}/Q_{\mathrm{net}}\}.$$

The above equation shows clearly the impact of changes in evaporation, integrated water vapour and net radiation on precipitation. If the vegetation is removed from a given region in the tropics, it reduces the evaporation from the surface. This will reduce the amount of water vapour in the atmosphere above. The removal of vegetation will also increase the albedo of the soil surface. In 1975, Jule Charney[9] had proposed that the impact of removal of vegetation will be primarily through the reduction in net radiation (caused by the increase in albedo of the soil surface). Moreover, a decrease in integrated water vapour will increase the OLR, hence causing a decrease in net radiation (in addition to that caused by increase in albedo). We find, therefore, that any decrease in water vapour content of the atmosphere, on account of decrease in vegetation, will lead to a decrease in precipitation through more than one mechanism. This is shown schematically in Figure 4. The effect of tropical deforestation on rainfall has been discussed recently by Zeng and

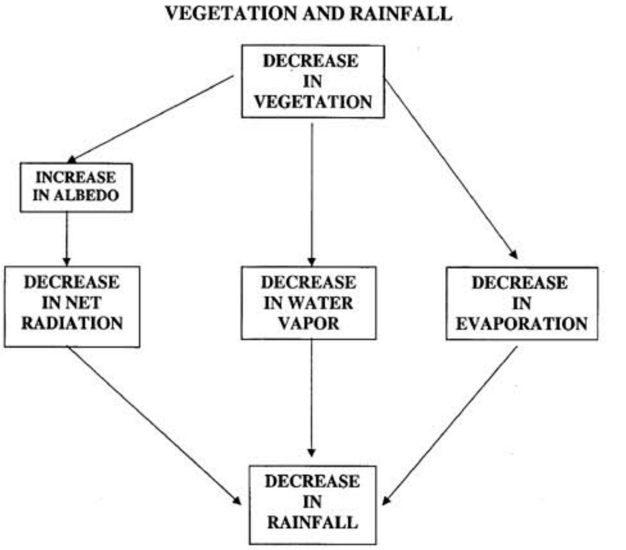

Figure 4. Schematic diagram for the interaction between vegetation and climate.

Neelin[10]. They assume, however, that the vertical stability of the atmosphere is invariant. The present model indicates, however, that the vertical stability of the atmosphere depends strongly upon the integrated water vapour and hence will not remain constant during deforestation.

We have demonstrated that it is possible to construct a simple thermodynamic model of the monsoon based on energy and moisture balance. This model is able to simulate accurately the seasonal variation of rainfall in large tropical continents. The model indicates that three important parameters control the seasonal variation of monsoon rainfall. They are the net radiation at the top of the atmosphere, evaporation and the vertically integrated water vapour in the atmosphere. The intriguing feature of this model is that it does not explicitly contain any parameter related to land–sea contrast in temperature. Hence it provides a new perspective on the factors that cause large seasonal variation in rainfall in the tropics. This model will be useful to understand why some GCMs are poor in simulating the seasonal variation of monsoon rainfall. The manner in which radiative and cloud processes are modelled in a GCM will determine the accuracy in the estimate of Q_{net}. The manner in which surface processes are parameterized will determine the accuracy in the estimate of surface evaporation. The manner in which the vertical transport of moisture is parameterized will determine the accuracy in the estimate of integrated water vapour (p_w). Moreover, the inaccuracy in estimate of one quantity (e.g. evaporation) may influence the accuracy of the other quantities (e.g. water vapour). The simple diagnostic model proposed here will be useful to identify which aspect of a GCM needs further refining.

1. Gates, W. L., *Bull. Am. Meteorol. Soc.*, 1992, **73**, 1962–1970.
2. Gadgil, S. and Sajani, S., *Climate Dyn.*, 1998, **14**, 659–689.
3. Neelin, J. D. and Held, I. M., *Mon. Weather Rev.*, 1987, **115**, 3–12.
4. Fortelius, C. and Holopainen, E., *J. Climate*, 1990, **3**, 646–660.
5. Kalnay, E. *et al.*, *Bull. Am. Meteorol. Soc.*, 1996, **113**, 2158–2172.
6. Barkstrom, B. R., Harrison, E., Smith, G., Green, R., Kibler, J., Cess, R. and ERBE Science Team, *Bull. Am. Meteorol. Soc.*, 1989, **70**, 1254–1262.
7. Xie, P. and Arkin, P. A., *Bull. Am. Meteorol. Soc.*, 1997, **78**, 2539–2558.
8. Li, C. and Yanai, M., *J. Climate*, 1996, **9**, 358–374.
9. Charney, J. G., *Q. J. R. Meteorol. Soc.*, 1975, **101**, 193–202.
10. Zeng, N. and Neelin, J. D., *J. Climate*, 1999, **12**, 857–872.

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