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One dimensional sensitivity studies with vegetation canopy

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सार – भारत में दो विभिन्न स्थानों पर भूमि की सतह पर वनस्पति विज्ञान के साथ एक आयामी अनुकरण के अध्ययनों से पता चलता है कि मृदा आर्द्रता, वनस्पति वाली एवं वनस्पति-रहित भूमि और वायुमंडल के मध्य प्रक्षुब्ध फ्लक्सो की दृढ़ता को नियमित करने में प्रभावी भूमिका होती है। वनस्पति के साथ, परिसीमा स्तर के अभिलक्षण, जैसे तापमान की प्रोफाइल, परिवर्तित हो जाते है। वनस्पति वितान द्वारा वायुमंडल के शुद्ध तापमान वनस्पति के विस्तार की मात्रा और आरंभिक अवस्थाओं पर निर्भर करता है। बोवन के अनुपात के परिकलित मान प्रेक्षणों के साथ मिलते हैं।

ABSTRACT. One dimensional simulation studies, with a vegetation canopy at the earth's surface at two different locations in India, show that the soil moisture has a dominant role in regulating the strengths of the turbulent fluxes between the atmosphere and the ground with or without vegetation. With vegetation, the boundary layer characteristics such as temperature profile get modified. The net warming of the atmosphere by the vegetation canopy depends on the amount of coverage and initial conditions. The calculated values of Bowen's ratio agree with observations.

1. Introduction

In many atmospheric simulation studies, the contribution of non-adiabatic turbulent fluxes to the atmosphere by the vegetation is usually neglected. The vegetation canopy forms a secondary source and sink of heat by the absorption and re-emission of incident solar radiation. Since the atmospheric variables are controlled by the exchange of heat and moisture with earth's surface, an explicit formulation accounting the above factors would be more appropriate. The objective of this study is to underline the importance of including vegetation effects in forecast studies and to evaluate the relative importance of the various parameters involved in modifying the boundary layer characteristics. Analytical solutions to the governing equations encompassing the complex structure of vegetation have not yet been developed. Thus, for simplicity, one dimensional numerical solutions are performed which can predict the response of atmospheric boundary layer (ABL) to large scale forcings.

2. A brief review of literature

Deardorff (1978) had shown that the ground surface temperature and its moisture content cannot be predicted with accuracy neglecting the overlying foliage layer. An one layer foliage parameteiization that extends continuously from the case of no shielding of the ground by vegetation to complete shielding was developed. However, the effects of stratification, influence of different types of canopy on the various parameters were not considered. Further, an application and validity of this scheme has not yet been tested for tropical regions.

The importance of vegetative surfaces in an arid region and over an urban area in the modification of climate are reported in literature (Otterman 1981, Terjung and O'Rourke 1981).

Including the vegetation and cloud shade effects simulations of winds over mountain slopes were obtained by Garrett (1983). The simulated values were found to be in poor agreement in complex terrain.

Yamada's (1982) model of the boundary layer structure over a flat terrain simulated nearly constant, low wind speeds within forest canopy with large wind shears near the top. However, Oke (1978) concluded that for a given wind speed, the atmosphere is more turbulent over a forest than any other natural surface, excluding the topographic effects. Yamada's model, based on complicated higher order closure scheme, is not applicable to complex terrains.

Although there are some existing foliage models (Paltridge 1970, Waggoner and Reifsnyder 1968) these

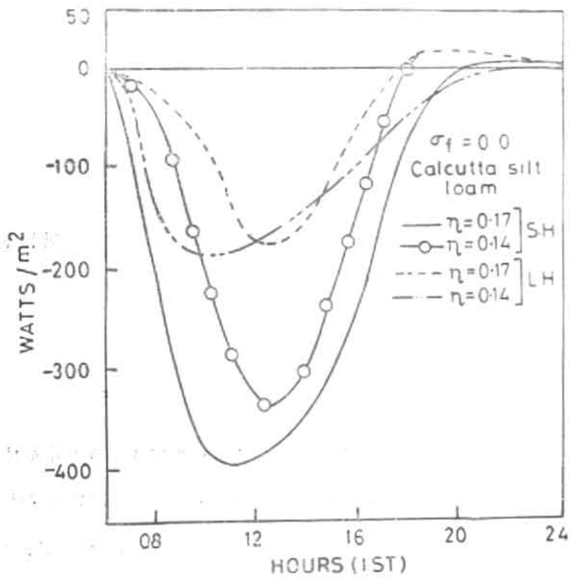

Fig. 1. Diurnal variation of diabatic fluxes with different intial soil moisture content (η) [unless specified, uniform η (cm³/cm³) profile is taken]

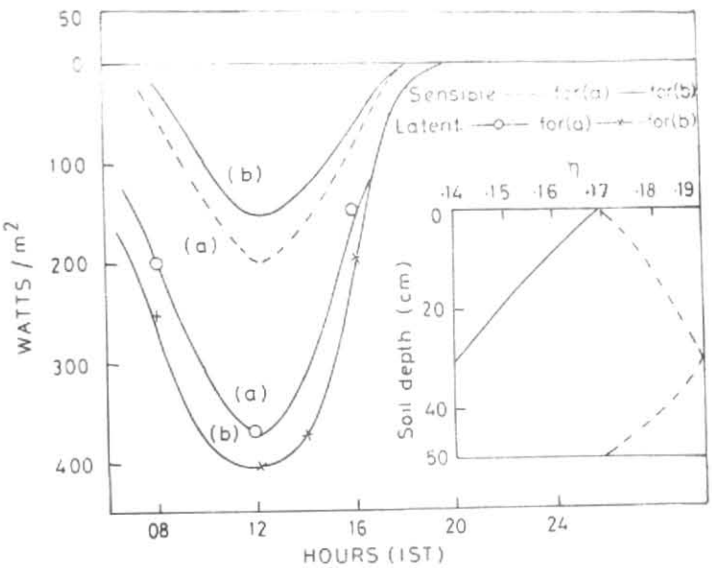

Fig. 2. Same as Fig. 1 but with different profiles of η : (a) linear and (b) parabolic (inset figure)

are multilayer and do not treat fluxes beneath the canopy nor allow canopy density to approach small values. The above mentioned models are one dimensional.

In this study, the atmospheric variables are predicted using the meso-scale model developed by Mahrer and Pielke (1977) by coupling with a soil parameterization scheme developed by McCumber and Pielke (1981) and Deardorff's vegetation parameterization scheme as modified by McCumber (1980). Sensitivity studies with the above formulations in extra-tropics have shown that the effect of vegetation is to produce milder climate near the ground. However, the above results are not applicable to tropics, since meso-scale circulations are the characteristics of the area studied besides the other factors such as soil type and moisture etc. Thus, a study relevant to tropics is necessary and appropriate. This will lead to a better understanding of the atmospheric boundary layer, which is essential for improved weather forecasts.

3. Description of vegetation parameterization scheme

In the modified vegetation parameterization scheme of McCumber (1980) the vegetation canopy is considered as a large leaf covering a fraction of the grid interval. The energy fluxes beneath the canopy and of the bare ground are calculated separately. The ground surface temperature (T_g) is obtained equating the sum of the incident and outgoing radiative fluxes to zero with

TABLE 1

Values of the soil parameters with dimensions taken from Table 11-5 pp. 395 : Pielke (1984)

Soil type	η_s (cm³/cm³)	ψ_s (cm)	$K\eta_s$ (cm/sec)	C_i (cal/cm°C)	λ (cal/cm² sec)
Silt loam (Calcutta)	.485	—78.6	7.2×10^{-4}	.304	5.3×10^{-4}
Sandy clay loam (Madras)	.420	—29.9	6.3×10^{-4}	.281	4.98×10^{-4}

where η_s=Porosity of soil,
$K\eta_s$=Saturated hydraulic conductivity and
C_i =Dry volumetric heat capacity of the soil

TABLE 2

Vegetation parameters used in simulation

	Grass	Tree
Emissivity (ϵ_f)	0.95	0.98
Albedo (α_f)	0.20	0.10
Stomatal resistance r_s (sec cm⁻¹)	4.0	8.0
Transfer coefficient (C_G)	0.0038	0.0176
Leaf area index (L_A)	$7\sigma_f$	$7\sigma_f$

TABLE 3

Initial potential temp. (θ), specific humidity (q) and soil temp. (T_s) profiles for Madras and Calcutta (00 GMT)

	Madras		Calcutta		Madras		Calcutta
Ht. (m)	θ (°K)	q (gm/gm)	θ (°K)	q (gm/gm)	Depth (cm)	T_s (°K)	T_s (°K)
Surf.	303.1	0.0199	301.3	0.0177	0.5	303.3	301.3
100	303.6	0.0188	302.1	0.0172	2	303.8	301.8
500	305.3	0.0161	306.2	0.0136	4	304.1	302.3
1000	307.7	0.0116	309.1	0.0098	7	304.7	302.8
2000	312.2	0.006	311.0	0.0062	12	305.5	302.8
3000	315.5	0.005	312.6	0.0045	17	306.3	302.8
4000	319.1	0.004	315.0	0.0033	22	307.0	302.8
5000	323.6	0.003	318.9	0.0019	27	308.3	302.3
6000	330.2	.0015	324.0	0.0010	37	307.5	302.3
					42	307.0	301.8

a proportional weightage to bare soil and the vegetation coverage as :

$$(1-\sigma_f)\{R_s(1-a_g)+\rho\, c_p\, u_*\,\theta_*+\rho\, L_{u*}\, q_*$$
$$+\epsilon_g(R_L-\sigma T_g^4)\}+\sigma_f(H_0+L_{E0}+R_N)$$
$$-\lambda\frac{\partial T_s}{\partial Z}=0 \qquad (1)$$

where,

λ thermal conductivity of soil

T_s temperature of soil layer

T_g temperature of ground surface

ϵ_g emissivity of the ground

a_g ground albedo

c_p Specific heat of air at constant pressure

ρ density of air

R_s incoming short wave radiation flux

R_L incoming long wave flux

L latent heat of evaporation

σ Stefan's constant

H_0 sensible heat flux below the canopy

L_{E0} latent heat flux below the canopy

R_N net longwave radiative flux of the canopy and the ground

q_* specific humidity

η volumetric moisture content (amount of moisture per unit volume of the soil)

In the above equation the first bracketed terms apply to the bare ground while the second group of terms, which appears with σ_f, refer to the fluxes beneath the canopy. The last term denotes the heat flux from the soil.

The foliage temperature is calculated using an energy budget equation similar to the Eqn. (1) shown above, assuming there is no canopy heat storage. The friction velocity, temperature and specific humidity (u_*, θ_*, q_*) are obtained with a weighted average for the bare soil and (σ_f) vegetation coverage fraction. Surface and foliage temperatures are obtained with Newton-Raphson iterative technique. The formulations for radiative fluxes, boundary layer parameterization, variations in the soil moisture content, soil albedo and the temperature etc can be seen in Mahrer and Pielke (1977) and McCumber and Pielke (1981), omitted here for brevity.

At the start of simulations the synoptic temperature, velocity and specific humidity profiles are specified;

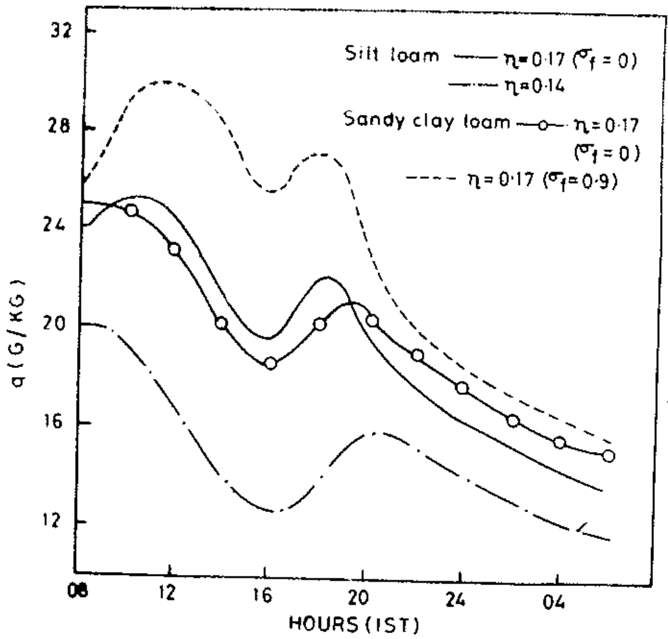

Fig. 3. A comparison of surface specific humidity (q) for different soil types and initial moisture contents for bare ground (σf=0) and with 90% of tree coverage (σf = 0.9)

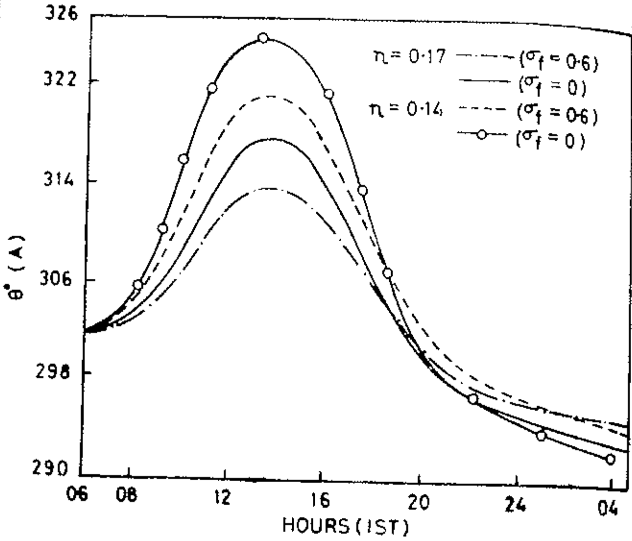

Fig. 4. Diurnal variation of surface temperature for silt with different η and σf

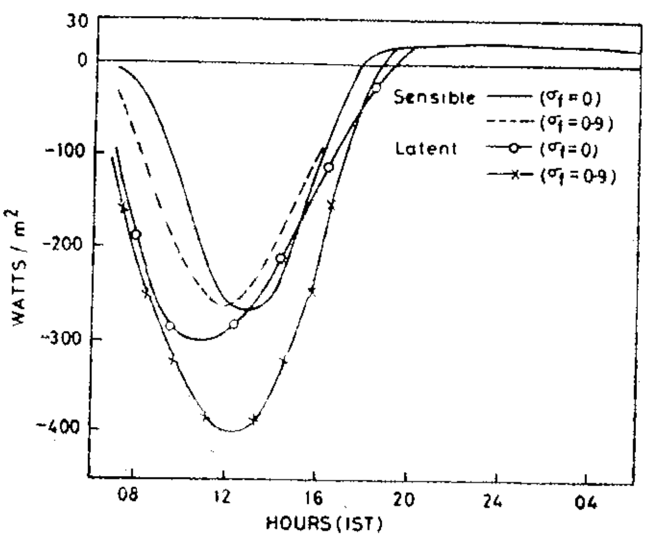

Fig. 5. The variation of sensible and latent heat fluxes for sandy clay loam soil (Madras) with η=.17 over bare ground and with trees σf=0.9

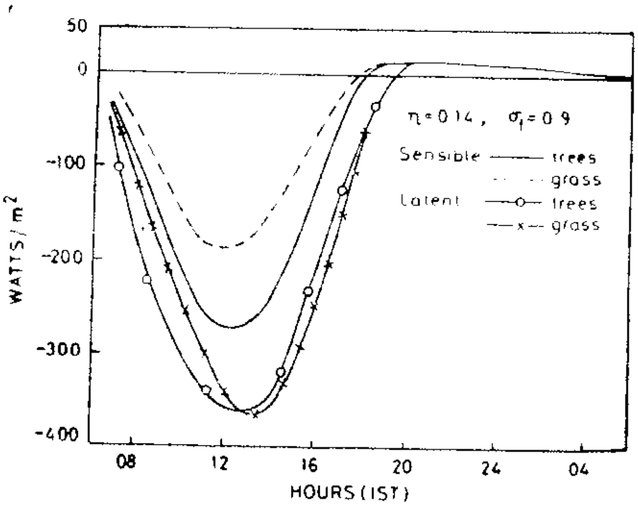

Fig. 6. Same as Fig. 5 with Calcutta data for η=.14, σf=0.9 and different types of foliage coverage

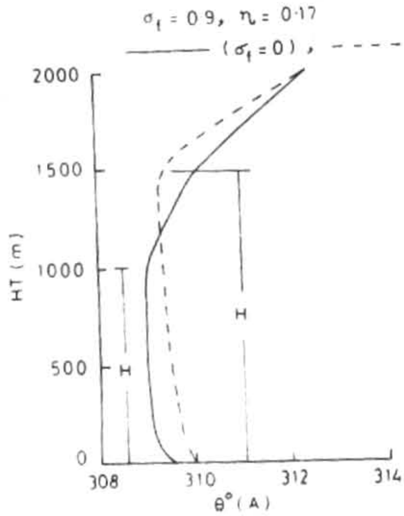

Fig. 7

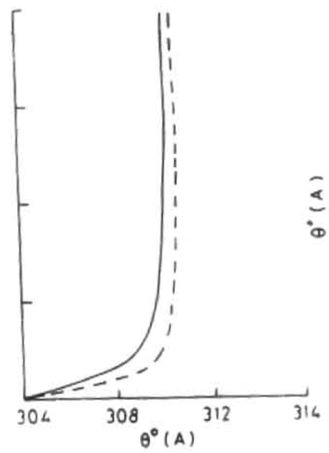

Fig. 8

Figs. 7 & 8. The potential temperature (θ) profiles using Madras data at 1400 LST (Fig. 7) and at 0600 LST (Fig. 8) over bare ground and tree covered surface ($\sigma f = 0.9$; $\eta = 0{\cdot}17$). The PBL heights are marked as H in the diagram

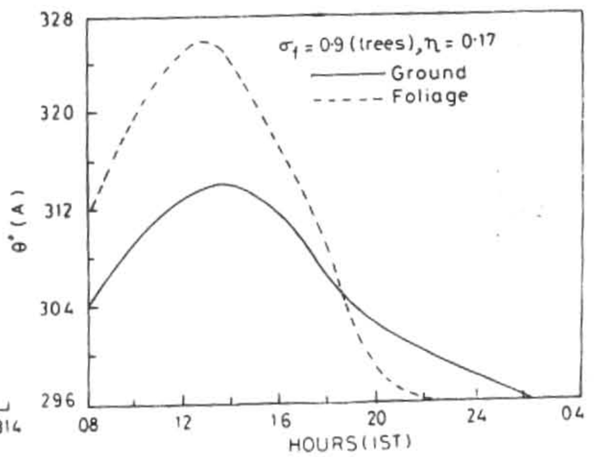

Fig. 9. A comparison of temperatures developed at the foliage and the ground with Madras data for $\sigma f = 0.9$ (trees), $\eta = 0.17$

usually a climatological mean of these variables are taken. In addition to these, the parameters that are specified for the simulations are (*i*) the surface drag coefficient (C_G), (*ii*) soil moisture profile (η), (*iii*) soil characteristics, (*iv*) vegetation coverage fraction (σ_f) and (*v*) vegetation characteristics. The above parameters have seasonal and diurnal variations, for example, the albedo and stomatal resistance of the vegetation depend on the solar elevation angle. In the absence of any detailed observations gross albedo values are used. Sensitivity tests with different values of stomatal resistance (r_s) and leaf area index (L_A), a variable which determines the sensible heat flux from the canopy, are made. The soil/vegetation types and the areal vegetation coverage fraction are obtained from atlas of agricultural resources of India. No distinction is made between the various types of grass and/or wheat/rice fields as the differences in their characteristics remain small. The temporal variations of the soil moisture and temperature are calculated *in situ*. The soil characteristics are given in Table 1 (Pielke 1984). C_G is calculated at the first grid level (Z) of the model above the vegetation canopy using the relation (Thom 1971) :

$$C_G = \left[\frac{K_0}{\ln\{(Z-d)/Z_0\}} \right]^2 \tag{2}$$

where K_0 is Von Karman's constant ($=0.35$), d is the displacement height and Z_0 is the turbulent roughness height. Deardorff's (1978) suggestions that $d=0.76\ H_C$, $Z_0 = 0.08\ H_C$, where H_C is the mean canopy height are adopted. These agree with Amazon forest observations. In our simulations, H_C is assumed to be 13 metres for trees and 1 metre for grasses. The first grid level (Z) is to be taken to be at 25 metres above the ground. The calculated values of C_G are given in Table 2. The remaining parametric values describing the type of vegetation are also listed in Table 2. Table 3 gives the initial synoptic temperature, humidity and soil temperature profiles used in our simulations for Calcutta and Madras. The soil temperature profiles were obtained from the climatological data at the respective simulated regions. All simulations were commenced at local sunrise time and continued till next day sunrise, *i.e.*, 24-hour simulations. The simulated results with various combinations of above parameters are presented.

4. Results and discussion

The influence of the initial soil moisture content on the intensities of turbulent heat fluxes for bare soil, with Calcutta synoptic data (Table 3), are shown in Fig. 1. In this and subsequent figures the outgoing fluxes are taken as negative and σ_f denotes the fractional coverage of land by vegetation ($\sigma_f = 0.0$ bare soil; $\sigma_f = 1.0$ complete coverage of land by vegetation). Fig. 1 shows that with large soil moisture the latent heat flux released to the atmosphere by evaporation is more than that of sensible heat flux. With drier soil ($\eta = 0.14$ cm^3/cm^3) the sensible heat flux exceeds the latent heat flux and attains its maximum around local noon, *i.e.*, at the time of maximum surface temperature. Maximum latent heat is released 2 hours ahead of the sensible heat flux around 1000 LST in the forenoon approximately at the time of maximum surface humidity.

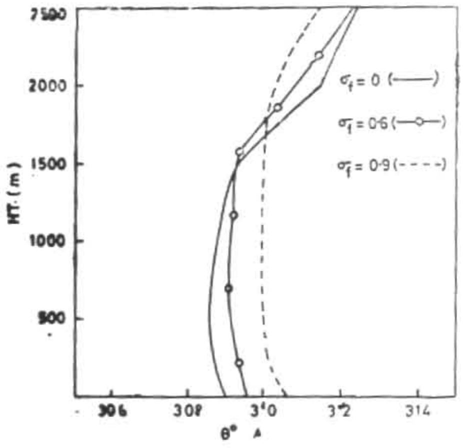

Fig. 10. θ profiles at 1400 LST for Calcutta (η=0.17) with σ_f=0, 0.6 and 0.9

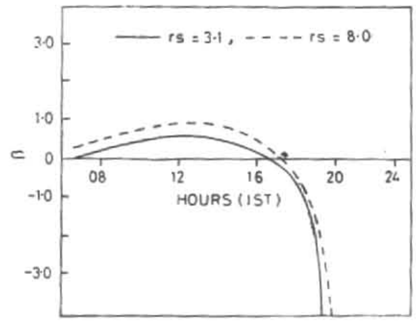

Fig. 11. Bowen's ratio (β) values at different hours for different resistance coefficients r_s with σ_f=0.9, r_s = 3.1 and 8.0

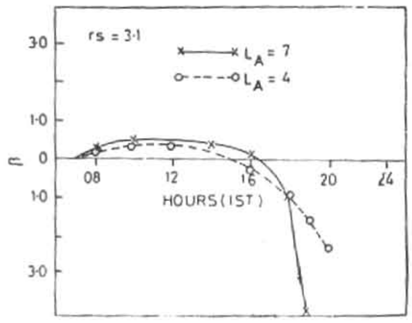

Fig. 12. Same as Fig. 11 with different leaf area index (L_A) values with r_s=3.1, L_A= 7 & 4

The above results are for an uniform distribution of moisture in the entire soil layer. However in nature, such uniform distributions are rarely found except in saturated conditions. To determine the influence of non-uniform moisture distribution in the soil layer on the intensity of turbulent fluxes, simulations with a linear decrease of moisture content with depth and with a parabolic distribution are made. These are shown in Fig. 2. The slight variations in the intensities are not significant, showing thereby, the moisture content in the top layer close to the ground surface is the primary factor involved in heat exchange rather than nature of its distribution within the soil layer.

The diurnal variations of the surface specific humidity and the surface temperature for bare soil (σ_f=0) are shown in Fig. 3 and Fig. 4 respectively. Surface humidity profiles exhibit a principal maximum around 1000 LST and a secondary maximum around local sunset or in early night hours. An analysis of the temperature profiles show that the increase in specific humidity around sunset time is due to a change in thermal stability of the atmosphere from unstable to inversion conditions. Such an increase in specific humidity with the onset of inversion has been recorded (Deabreusa *et al.* 1988). With initial moisture content (η=0.17 cm^3/cm^3) the surface specific humidity at all hours is higher than with relatively dry soil, shown in the above figure.

The diurnal variation of surface temperatures (Fig. 4) show that with drier soil higher temperatures 8° C are developed in comparison to the wet soil. While the temporal variations are fairly symmetric irrespective of the soil moisture content, the rate of heating and cooling with drier soil is more than that of wet soil, a feature of common knowledge.

Fig. 5 compares the magnitudes of heat fluxes obtained with Madras synoptic data for sandy clay loam soil, with no vegetation and with an initial soil moisture η = 0.17 cm^3/cm^3. The variations in the magnitudes between different soil types can be seen by comparing the results of Fig. 1 with Fig. 5. For the same initial moisture content, the difference between the sensible and the latent heat flux with sandy clay loam soil is of the order of 50 W/m^2; with slit loam soil the difference is of the order of 200 W/m^2. The absolute magnitudes of the individual fluxes are not compared due to latitudinal differences (Madras 13°N, Calcutta 23°N) between the simulated regions.

The above results, not reported in Indian literature, show the intensity of turbulent heat fluxes are a function of soil type, soil moisture and the geographical location of the simulated region. Under the same climatic conditions, the net heat exchange to the atmosphere, with no vegetation, depends principally on its soil moisture content (η) rather than its distribution in the soil layer, which agrees with McCumber and Pielke (1981) results.

4.1. *Influence of vegetation* — The vegetation canopy modifies the boundary layer climate in two ways: (1) by cooling the ground surface beneath it (shadow effect) & (2) as a secondary source and sink by the re-emission and absorption of incident solar radiation.

For this study the types of vegetation were taken to be that of trees and wheat/rice field or grasses whose parametric values, specified at the start of simulations, are given in Table 2.

The diurnal variation of surface temperature with tree canopy of fractional coverage σ_f=0.6 and different initial moisture contents are shown in Fig. 4 along with bare soil results. The reduction in the maximum surface temperatures are of the order of 5°C to 3°C and the intensity of reduction is more with drier soil (η=0.14 cm^3/cm^3).

A similar result, not shown here, is also obtained with Madras data. It is clear that the moisture content in the soil plays a vital role in determining surface temperature variation with or without vegetation. The increase in the intensities of turbulent heat fluxes in the presence of vegetation are shown in Fig. 5 along with bare soil results. With a denser vegetation coverage ($\sigma_f = 0.9$) the intensities increase, which implies in dense forests, the soil characteristics play a minor role in determining the boundary layer climate. A proportionate reduction in the magnitudes of turbulent heat fluxes cannot be expected since the latent heat flux is an implicit function of surface temperature whereas the sensible heat flux is directly proportional to surface and foliage temperatures.

A comparison of surface specific humidities with different amount of vegetation coverage ($\sigma_f = 0.0$ and $\sigma_f = 0.9$) are shown in Fig. 3. Since the foliage canopy regulates the evaporation of surface moisture to the atmosphere, higher specific humidities are obtained throughout the simulation period with a denser coverage.

The relative importance of vegetation type, trees or grasses, on the heat exchange to the atmosphere is shown in Fig. 6. It is clear a tree canopy is more effective in heat transport to the atmosphere than a grass field under identical conditions due to lower albedo and high stomatal resistance of trees, in agreement with (Oke 1978) conclusion. However, this result is in contradiction with McCumber's (1980) conclusion that the vegetation effect is independent of the type of vegetation. This may be due to the vegetation types such as coniferous trees in mid-latitudes.

The cumulative effect of vegetation coverage and the surface moisture on the boundary layer temperature profiles are shown in Figs. 7 and 8. A net warming of the atmosphere by about 0.5°C in mid-day temperature profile (Fig. 7) and a slight warming (0.1°C) in early morning profile (Fig. 8) are developed with vegetation. The temperature profile obtained with bare soil are shown in thick lines in the above figures for comparison. The boundary layer depth in response to net warming increases by about a few hundred metres as seen from the above Fig. 7. The net warming and subsequent increase in PBL depth shows an increase in the intensity of turbulence in the PBL. The warming in day time is combined effect of foliage albedo and the re-emission of long wave radiation by the plant canopy which develops a higher temperature than the ground as shown by the diurnal variation curve in Fig. 9. The simulated potential temperature profiles at 1400 LST for a forested surface, with different amount of vegetation coverage are compared in Fig. 10 using Calcutta data. The net warming is proportional to the vegetation coverage fraction.

The hourly Bowen ratio (β) values obtained with two different specifications of stomatal resistance (r_s) and leaf area index (L_A). for tree coverage, are shown in Figs. 11 and 12. Higher stomatal resistance yield higher β values during day light hours while it remains insensitive with the change in L_A values (Fig. 12). In both the diagrams β values remain less than unity from sunrise to sunset. $\beta < 1$ indicates climate is likely to be warm and highly moist. The above β values agree with Amazon forest observations, where β varies between 0.05 & 0.85 during day time (Deabreusa *et al.* 1988). Around sunset time and during night hours negative $\beta(> -3.0)$ are obtained due to change in thermal stability of atmosphere and a subsequent change in the direction of sensible heat flux (from the surface to the surface). The decrease of evaporative flux from the ground during night hours with the increase of L_A yield larger negative β values. However, the absolute magnitudes of these fluxes remain an order of magnitude less than that obtained during day time.

The presented results are in qualitative agreement with other studies. However, for example, the net warming in day time in our study is much smaller than that obtained in McCumber (1980) simulations. Since similar results are obtained with silt loam soil, the differences in magnitude are due to initial synoptic temperature and moisture profiles used in our simulations.

This study has shown the relative importance of soil moisture, soil type, vegetation type, coverage and its parameters with respect to the modification of boundary layer characteristics such as its depth, temperature structure and humidity. These are to be verified by appropriate meteorological observations. Future atmospheric prediction studies should include the vegetation effects into their models in view of the gross errors which can occur when the overlying foliage layer is ignored.

5. Conclusion

The sensitivity tests with different initial moisture contents and soil types used in our study show the soil moisture at the surface determines the intensities of turbulent heat fluxes between ground and atmosphere. The increase in surface specific humidity at the sunset time is determined by the atmospheric conditions such as the formation of inversion layer rather than by the nature of earth's surface. With vegetative cover a net warming of the atmosphere and subsequent growth in PBL depth in day proportional to the vegetation coverage are obtained.

A higher degree of turbulence can be expected over a tree covered surface rather than a grass canopy due to increased heat transfer under the same synoptic conditions. The simulated β values agree with the reported forest observations.

Further studies should include the effects of advection, radiative flux changes by clouds, surface/foliage cooling by rainfall, roughness discontinuities with a generalisation to higher dimensions. These are currently in progress.

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