PRECIPITATION SCAVENGING AND ATMOSPHERE-SURFACE EXCHANGE

Volume 2—The Semonin Volume: Atmosphere–Surface Exchange Processes

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Advective Influences in the Amazonian Forest Terrain

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ABSTRACT

This study was conducted to evaluate the performance of the energy balance components at a Ducke Reserve Forest (DRF) site ($2^{o}57'$ S; $59^{o}57'$ W) in the Amazon Basin, a region where a significant proportion of the energy consumed by evapotranspiration is supplied from advective sensible heat. Analyses suggest that the values of γ (i.e., the ratio of eddy diffusivity coefficients of heat K_H and water vapor K_W) vary from -2.67 to 2.70 in the Richardson number range -0.17 to 0.28 and the negative sign of γ arises in the cases of countergradient heat flux. There is, however, evidence in the literature to support an assumption that $\gamma > 1$ can exist under stable conditions.

1. INTRODUCTION

It is commonly accepted that humid tropical regions have a high radiation income and abundant precipitation so that the air is very humid and hot. It is easy to conclude that these climatic conditions generate exuberant tropical vegetation. It has long been recognized that the atmospheric and soil-vegetation systems are dynamically coupled through the physical processes which bring about the transport of thermal energy and water mass across the land surface. This phenomenon explains the importance of current knowledge regarding the micrometeorology of forests. A description of the characteristics of atmospheric turbulence in the surface layer over forest canopies is essential for the understanding of the micrometeorological environment of forests. Although measurements of mean flow and turbulent fluctuations had been made in this layer (e.g., Raupach and Thom, 1981; Shuttleworth et al., 1982; Moore, 1976; Viswanadham et al., 1987, 1990; and many others), the turbulence aspects including the advective conditions have not yet been studied sufficiently.

It is very difficult to calculate the advection in a tropical rain forest (Amazon forest) because some of the advective terms are not yet measurable due to the great heterogeneity of species and tree sizes. Nevertheless, Warhaft (1976) provided expressions for the vertical heat and moisture fluxes in the stratified boundary layer using two different models (Donaldson 1972; Launder 1975) for the pressure strain term in the equations for the conservation of heat and moisture flux. The parameterization of Warhaft is only strictly applicable to homogeneous turbulence and we expect, at best, a qualitative view of the situation near ground. Warhaft predicted that the eddy transfer coefficient for sensible heat (K_H) would be less than the eddy transfer coefficient for latent heat (K_W) when the sensible heat flux (Q_H) was directed downwards and the water vapor flux (Q_E) was upwards, which are the conditions within an advective inversion (Warhaft, 1976). A second possible reason is that in turbulent transport, K_H and K_W depend on the characteristics of the whole flow and not simply on local properties. Thus, we expect K_H and K_W to depend on the respective temperature and humidity profile shapes and more importantly on the vertical turbulence statistical properties i.e., the vertical inhomogeneity, skewness and time scale of the turbulence, and also on the source distribution (Raupach, 1987 and Weil, 1990); if these are different, then differences between K_H and K_W are possible. The dependence of K_H on the turbulence properties and source location (bottom or top) in a convective boundary layer has been shown by Weil (1990). For example, transport within forest canopies is dominated by infrequent, large eddies, which bring air down from above the canopy and generates a negative vertical velocity skewness, where the temperature is similar to that within the canopy but humidity is lower. In these circumstances, simple K-theory is invalid and counter-gradient fluxes are observed (Deardorff, 1966; Denmead and Bradley, 1985; Raupach, 1987). A similar situation is possible within an advection inversion where the temperature is a maximum at some height but the humidity profile is monotonic (i.e., having little or lack of variation).

The present work describes the field experiments at a Ducke Reserve Forest (DRF) site (2°57'S, 59°57'W) in the Amazonian evergreen forest, Amazonas, Brazil. The main objective was to present energy fluxes under advective conditions and, to establish the relationship between eddy transfer coefficients from direct measurements of eddy fluxes of Q_H and Q_E together with simultaneous measurements of the profiles of temperature and humidity.

2. BASIC RELATIONSHIPS

The flow over a forest canopy of height h (m) exerts an influence throughout a roughness sublayer that extends considerably above height z (m) = h. At its upper limit, the roughness sublayer merges with an inertial sublayer in which the Monin-Obukhov's similarity theory can be applied. Generally, this similarity theory applies to horizontal homogeneous conditions. However, the Amazon advective site in DRF is a horizontal inhomogeneous one.

According to the constant flux layer similarity theory (Monin and Yaglom, 1977), the eddy transfer coefficients for sensible (K_H m^2 s^{-1}) and latent (K_W m^2 s^{-1}) heats may be written in terms of the fluxes and the dimensionless profile gradients as

$$K_H = -Q_H/[\rho C_p(\partial\theta / \partial z)] = ku_*(z-d)/\phi_H \quad (1)$$

and

$$K_W = -Q_E/[\rho\lambda(\partial q / \partial z)] = ku_*(z-d)/\phi_W \quad (2)$$

Also, the fluxes of momentum (τ kg m^{-1} s^{-2}), sensible(Q_H W m^{-2}) and latent (Q_E W m^{-2}) heats in terms of eddy covariances,

$$\tau = -\rho\overline{u'w'} \text{ or } u_* = (|\tau/\rho|)^{1/2} \quad (3)$$

$$Q_H = \rho C_p \overline{w'\theta'} \quad (4)$$

$$Q_E = \rho\lambda \overline{w'q'} \quad (5)$$

where ρ is the density of air (kg m^{-3}); k is von Kármán constant (= 0.4); u_* is the friction velocity (m s^{-1}); θ is the mean potential temperature (K); q is the specific humidity (kg kg^{-1}); C_p is the specific heat of air at constant pressure (J kg^{-1} K^{-1}); λ is the latent heat of vaporization of water (J kg^{-1}), d is the zero-plane displacement (m); ϕ_H and ϕ_W are the nondimensional gradient functions for heat and water vapor, respectively and u', w', θ' and q' are departures from means of the horizontal wind velocity, vertical wind velocity, potential temperature and specific humidity, respectively, and overbars denote time averages.

If we define Bowen ratios from fluxes (β_f) and gradients (β_g) as

$$\beta_f = (C_p/\lambda) (\overline{w'\theta'}/ \overline{w'q'}) \quad (6)$$

and

$$\beta_g = (C_p/\lambda) (\Delta\theta/\Delta q) \quad (7)$$

where $\Delta\theta$ and Δq are measured between the same vertical heights, and finite differences have been substituted for differentials, then from Equations (1) and (2)

$$\Upsilon = (K_H/K_W) = (\beta_f/\beta_g) \quad (8)$$

There are theoretical reasons why K_H may not always equal K_W. For example, buoyancy may act differently on the heat-carrying and vapor-carrying parcels of air, but this is only possible if these parcels are at least partially distinct (e.g. Launder, 1978).

The Richardson gradient number (Ri) is defined as

$$Ri = [(g/\bar{\theta})(\partial\theta / \partial z)]/(\partial u / \partial z)^2 \qquad (9)$$

where g is the acceleration due to gravity (m s^{-2}) and $\bar{\theta}$ is the average potential temperature of the layer under consideration. The above equations can be utilized to compute ratios of eddy transfer coefficients and Ri from simultaneous measurements of fluxes and profiles over the Amazon forest.

3. EXPERIMENTAL SITE AND MEASUREMENTS

The place selected for the experiment is in the tropical evergreen forests of the Amazonas Basin, Brazil. The measurements were made using a 45 m scaffolding tower at a site (2°57'S; 59°57'W), situated in the Ducke Reserve Forest (DRF), 26 km from Torquato Highway, Manaus, Amazonas. The altitude of the DRF site is 84 m above mean sea level. The DRF belongs to the Instituto Nacional de Pesquisas da Amazônia (INPA), Manaus, Amazonas, Brazil. Viswanadham et al. (1990) have presented details about the experimental site and measurements of various necessary parameters at the DRF tower.

4. RESULTS

This work describes some preliminary results of the advective conditions from the Amazonian rain forest micrometeorological experiment. Several intensive periods of data collection from micrometeorological instrumentation were planned from 1983 to the next few years. Already data have been collected in seven experiments. The third (13th July to 31st August, 1984) and fourth (15th March to 15th May, 1985) expeditions data sets were used to obtain various results during advective conditions. Stable conditions over the Amazon forest canopy usually occur during early morning, afternoon and night hours, when available energy (i.e., net radiation Q^*) at the canopy surface is relatively low, so that sensible and latent heat fluxes are small and thus relatively difficult to measure. However, local advection, arising from a transition from a dry to a wet surface, typically causes locally stable conditions accompanied by very large fluxes. In other words, in an environment with warm air above colder air, turbulence moves warm air down the gradient to cooler air, i.e., a downward or negative flux. One criterion for the identification of advective influences is that Q_E/Q^* is consistently greater than unity on an hourly basis (Oke, 1979; Viswanadham and André, 1983).

The wind tunnel results of Raupach et al. (1980) show that over rough closed canopies, the logarithmic wind profile law applies only above $h + 1.5\ell$, where h (= 35 m) and ℓ (= 2 m) are the height and the transverse length scale of the roughness elements, respectively. Viswanadham et al. (1990) have obtained that the value of $h + 1.5\ell$ is about 38 m for

DRF site. So, measurements below 39 m were not used to compute the roughness length z_o, the zero plane displacement d, and gradients from the profiles. The iterative procedure of Robinson (1962) is used to determine the roughness length z_o (m) and d from 80 quasi-adiabatic wind profiles of the Amazon forest. The mean values of 3 m and 28 m are obtained for z_o and d respectively. Various hourly data sets were selected using the above criterion for advection situations. Equations (1-2) or Equations (6-8) are chosen to compute the ratio γ with the availability of the appropriate data sets for the layer 41.04 to 44.66 m above the surface. The stability parameter Ri for the same layer is obtained from Equation (9). The parameter Q^*, an important climatic characteristic, represents the amount of energy available to atmosphere and canopy processes (Figures 1-4).

Figure 1 presents fluxes of Q^*, Q_H, Q_E, and the residue term RE for a fine day in August 1984. There is a fairly constant partition of the available energy during daylight hours as latent and sensible heat, often some sustained evaporation into the early evening, and outgoing radiation of the order of 40 to 50 W m^{-2} as the forest cools during the night. The suppression of turbulent interaction between the forest and the atmosphere at night, typified by the low energy fluxes, is a feature of the data and applies to all the turbulence variables. The data in Figure 1 do not satisfy energy balance at night. This may be due to an overestimation of positive Q_E and an underestimation of negative Q_H by Hydra (a battery-powered eddy correlation flux measuring device) during night hours. In fact, the detailed behavior of measured turbulent fluxes at night is difficult to interpret, and it is suspected that hourly measurements produced by the Hydra are subject to apparently haphazard fluctuations of the order of 10 W m^{-2} (see Viswanadham et al., 1990 for discussion of Hydra measurements). This may lead to the apparent lack of energy balance closure at night in Figure 1. The term RE is subject to much erratic variation during the day period. It is assumed that the physical heat storage (i.e., soil plus canopy storage heat fluxes) is equal to RE after neglecting the biochemical heat storage due to plant photosynthesis. Heat storage within the forest canopy can be a significant portion of the energy budget for forest stands, and this is apparent in the figure. There is a loss of energy to storage in the morning as temperatures rise, and this condition reappears in the late afternoon. From Figure 1, it is evident that Q_E exceeds the net radiation before 0700 hours and after 1600 hours, and Q_H is directed toward the forest canopy. During this period the Bowen ratio β_f ($=Q_H/Q_E$) is negative and the ratio Q_E/Q^* is greater than unity for a period of 3 or 4 hours after 1600 hours. In nonhomogeneous terrain like the Amazonian forest, advection of hotter and/or drier air from upwind areas helps to increase Q_E rates. The advection of drier or warmer air will also be observed with an increase of vapor pressure deficit. Sensible heat Q_H has been drawn from the air and consumed in evaporation during advection conditions. The well-defined subsidence inversions (see de Bruin, 1983; Sá et al., 1988) and the absolute values of the time rate of change of the vapor pressure deficits of the order 0.5 to 2 mbar h^{-1} were observed above the canopy. All these were responsible to the results (i.e., $Q_E > Q^*$) presented in Figure 1 for morning hours (i.e., before 0700 hours) and afternoon hours (i.e., 1600-2000 hours) over the Amazonian forest surface.

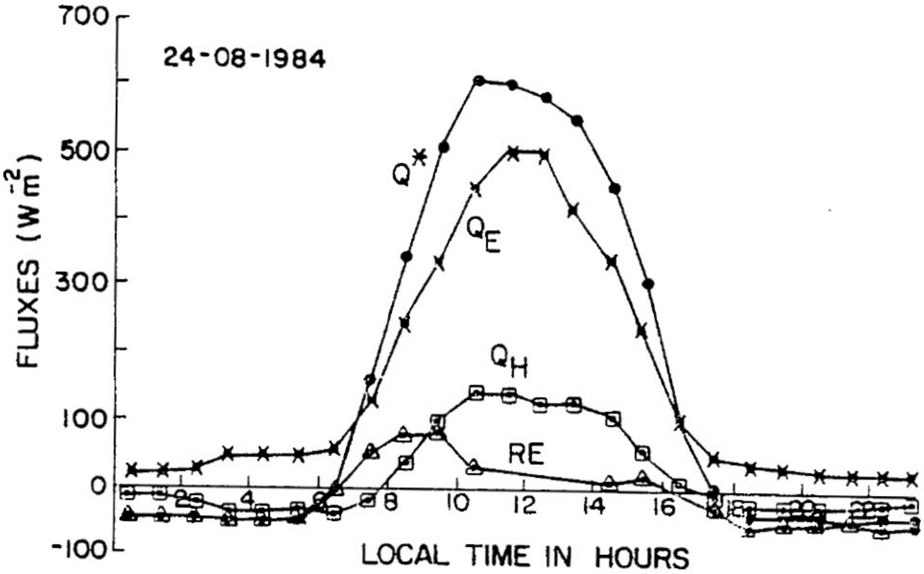

FIGURE 1. Diurnal variation of the instant energy balance components above the Amazonian rain forest during August 24, 1984.

The selected advective condition results are presented in Figure 2 which suggests good agreement between the eddy heat fluxes ($Q_H + Q_E$) and the available energy (Q^*) with a high correlation coefficient of r = 0.96. Agreement is typically better than 10%, which gives us confidence in the accuracy of the flux measurements.

The advective conditions over the Amazon forest were generally observed during the morning and evening hours. They were also noticed after the passage of local convective showers (i.e., before and after noon hours). In Figure 3, the Bowen ratio (β_f) values as a function of stability parameter Ri for these conditions are presented. Sometimes these advective conditions were also observed in quasi-neutral (i.e., $-0.02 \leq Ri \leq 0.015$) stabilities (see Crawford, 1965). In such situations, the positive Q_H and Q_E values were recorded and consequently β_f is positive in

Figure 3. Negative β_f values merely indicate that the two fluxes have different signs. Figure 3 shows that this is common in advective conditions when the sensible heat flux is downward (negative), but evaporation continues so that Q_E is away from the canopy surface (positive). Actually, this was due to the importance of night radiation cooling which produced cloud fog above the forest vegetation and water condensation on the forest foliage. The importance of dew in humid tropical forest like the Amazon is well known. In the early morning hours, fog lifted and the water vapor pressure was higher than closer to the foliage surface where water vapor had condensed. This gave the impression that water vapor was advected to our system, giving positive values of Q_E.

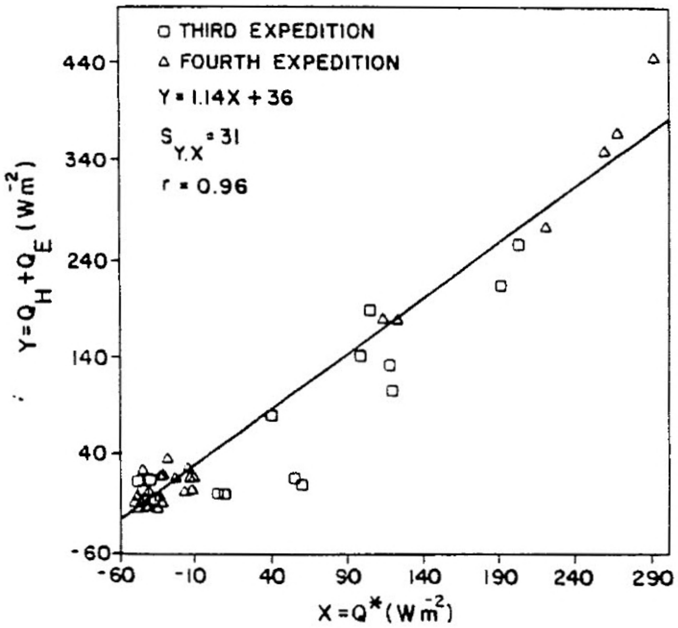

FIGURE 2. Comparison of sum of sensible and latent fluxes (Q_H + Q_E) measured above the Amazon forest by eddy correlation with the available energy (Q^*) under advective conditions.

In Figure 4, the ratio Υ as a function of Ri is presented for the advective situations. It shows that Υ values vary from -2 67 to 2.70 in the stability range -0.17 to 0.28. Specifically, downward fluxes of Q_H were accompanied by negative gradients of specific humidity. So, in Figure 4 the ratio Υ is positive in the advective conditions (see Equations 6 and 7). On the other hand, the negative Υ is generally associated with negative sensible heat fluxes and positive gradients of temperature and humidity. Thus, the best documented examples of counter-gradient flow are those for sensible and latent heat fluxes. The effects of these inconsistencies (i.e., $\pm\gamma$ values) on the calculation of scalar transport above the canopy from gradients can be seen in Figures 3 and 4, which compare the directly measured heat and water vapor fluxes above the forest with those which would be calculated from the gradients of temperature and humidity via Equations (1) to (8).

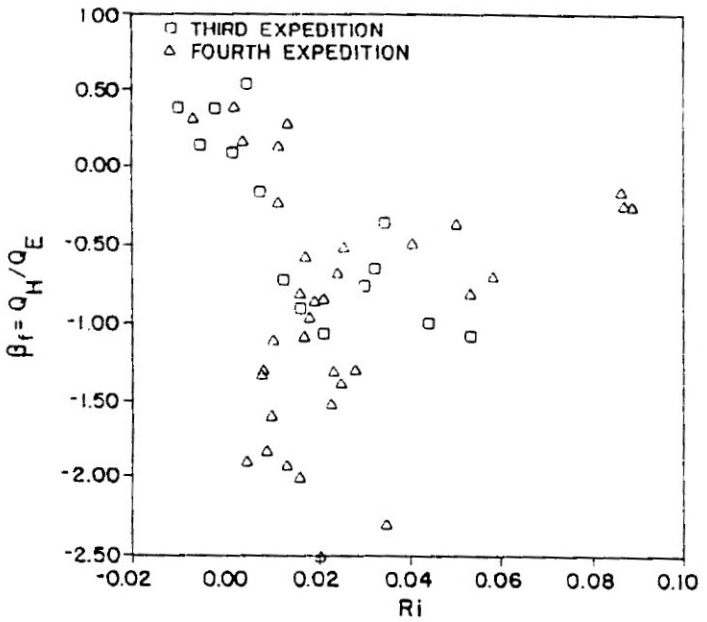

FIGURE 3. The Bowen ratio β_f (= Q_H/Q_E) as a function of stability parameter Ri for advective conditions.

The present results indicate that γ is greater than unity under stable conditions and also some negative values are noticed for γ (Figure 4). Literature on the subject of transfer coefficient relationships offers many diverse discussions. Rider (1954) noted approximate equality of the exchange coefficients regardless of stability conditions in most cases but exceptions were found in which the diffusivity for heat was much larger than that for vapor. At Ri = 0.004 the values of γ were about 1.18 and 1.09 for his observation numbers 44 and 46, respectively. Pruitt and Aston (1961) concluded that γ was very nearly 1.0 in unstable conditions, about 1.2-1.3 at Ri=0, and as high as 2-3 under very stable conditions. Estimates of the diffusivity ratios have been given by Munn (1966) based on data from Project Green Glow at Hanford, Washington and from

O'Neill, Nebraska. The values were not obtained from fundamental eddy correlations but were derived from an indirect use of the energy balance equation. From these methods Munn (1966) presented that the values of γ during night time are 1.13 and 1.71 for Green Glow and O'Neill, respectively. Campbell (1973) concluded that over the stable range $0 < Ri < 0.5$ the ratio $\gamma \geq 1$. For strongly stable case of $Ri \cong 0.5$, K_H was sufficiently greater than K_W to lead to errors of greater than 10% in the calculations of Bowen ratio and the corresponding Q_E .

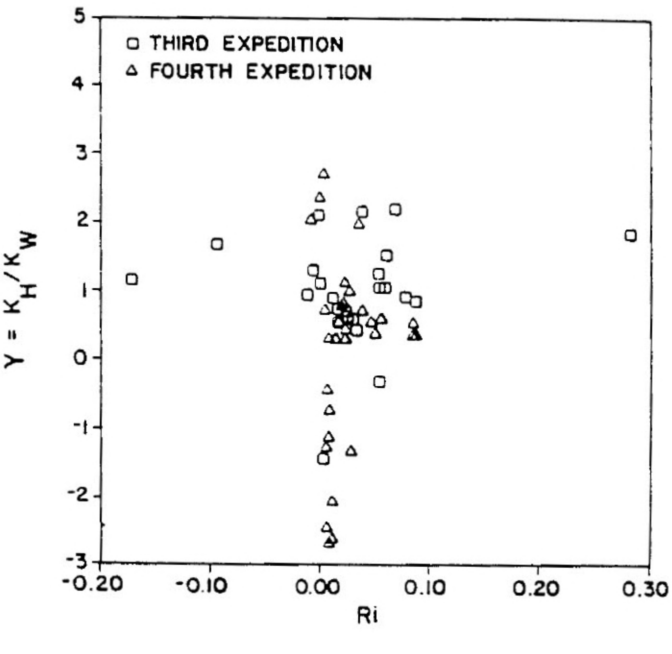

FIGURE 4. The eddy transfer coefficients ratio γ ($=K_H/K_W$) as a function of stability parameter Ri during advective conditions.

A number of experiments done in Nebraska with irrigated crops bounded by dry areas have led to comparisons of K_H and K_W. Results of Verma et al. (1978) based upon lysimetry, and Motha et al. (1979), from eddy covariances, gave γ of about 1.5. The analysis of Verma et al. may be questionable (Hicks and Everett, 1979; Verma et al., 1978). The work of Motha et al. (1979) should be more reliable since the eddy flux, obtained with a Gill propeller and fine wet and dry bulb thermometers, were measured separately and at the same height as the gradients. Under advective conditions, there are vertical divergences of the separate fluxes. Lang et al. (1983) obtained conflicting results for similar circumstances. With direct flux measurements, they concluded that γ approached unity near neutrality and decreased to about 0.65 with increasing stability.

The above evidences suggest that $\gamma = 1$ or $\gamma < 1$. There is, however, support for the condition that $\gamma > 1$ under stable conditions. Viswanadham et al. (1987) presented a critical discussion about $\gamma > 1$ under stable conditions. The present results are consistent with Raupach (1979) and Viswanadham et al. (1987). The most dramatic examples of negative values of γ associated with counter-gradient heat and water vapor fluxes over the real forest canopies. Even after applying corrections for the eddy correlations flux measurements, $\gamma > 1$, which conflicts with theoretical predictions of Warhaft (1976).

The discrepancy has yet to be explained. It may arise from a failure to account for all the terms in the Reynolds stress equations, from differences in the physical locations of the sources and sinks for momentum, heat and water vapor within the forest, or, perhaps, from differences in the modes of transfer of momentum and scalars at foliage surfaces (Denmead and Bradley, 1985). These results under advective conditions are important for mesoscale models which illustrate the links among the radiation budget, the climate and the circulation of the atmosphere. It was thus clear that equivalence of K_H and K_W under stable conditions should not be taken for granted and that relationships between transfer coefficients under an advective inversion need further investigation. Limited space prevents further discussions of our results.

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DISCUSSION

J. C. WEIL. K_H and K_W depend not on the temperature and humidity profiles but more importantly on the vertical turbulence statistical properties.....i.e., the vertical inhomogeneity, skewness, and time scale.....and also on the source distribution. The dependence of K_H on the turbulence properties and source location (bottom or top) in a convective boundary layer has been shown by Weil (1990) and Wyngaard and Weil (1991).

Y. VISWANADHAM. This information included in the text.

J. C. WEIL. In essence, you state that the countergradient vapor flux causes the negative Υ values. Is this mainly because the canopy is transpiring and produces a positive humidity gradient?

Y. VISWANADHAM. Yes.