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TO: 910/Chief, Laboratory for Atmospheres

FROM: 910/Laboratory for Atmospheres/Sounder Research Team

SUBJECT: Paper "Utilization of Satellite Data in Land Surface Hydrology: Sensitivity and Assimilation"

It is appropriate for Dr. Lakshmi, who is an employee of SAIC supporting the Sounder Research Team, to be the lead author for the paper "Utilization of Satellite Data in Land Surface Hydrology: Sensitivity and Assimilation" which we are submitting to the Journal of Hydrological Processes for publication. The research done for this paper is basically his. This work deals with the investigation of the sensitivity of potential evapotranspiration to air temperature and vapor pressure as well as a scheme to assimilate land surface temperature data in a hydrological model. Lastly, the cumulative errors in potential evapotranspiration arising from using satellite derived values of air temperature and vapor pressure rather than field observations are computed.

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Enclosure
Paper w/Approval Form

Abstract

This paper investigates the sensitivity of potential evapotranspiration to input meteorological variables, viz- surface air temperature and surface vapor pressure. The sensitivity studies have been carried out for a wide range of land surface variables such as wind speed, leaf area index and surface temperatures. Errors in the surface air temperature and surface vapor pressure result in errors of different signs in the computed potential evapotranspiration. This result has implications for use of estimated values from satellite data or analysis of surface air temperature and surface vapor pressure in large scale hydrological modeling. The comparison of cumulative potential evapotranspiration estimates using ground observations and satellite observations over Manhattan, Kansas for a period of several months shows very little difference between the two. The cumulative differences between the ground based and satellite based estimates of potential evapotranspiration amounted to less that $20mm$ over a 18 month period and a percentage difference of 15%. The use of satellite estimates of surface skin temperature in hydrological modeling to update the soil moisture using a physical adjustment concept is studied in detail including the extent of changes in soil moisture resulting from the assimilation of surface skin temperature. The soil moisture of the surface layer is adjusted by $0.9mm$ over a 10 day period as a result of a $3K$ difference between the predicted and the observed surface temperature. This is a considerable amount given the fact that the top layer can hold only $5mm$ of water.

Keywords: hydrological modeling, satellite data, sensitivity, assimilation

1 Introduction

The sensitivity of evapotranspiration to various land and atmospheric variables is important in order to identify variables critical to the land surface water balance. These critical variables need to be modeled and measured with care and accuracy if we wish to obtain accurate estimates of evapotranspiration and land surface water budgets.

The sensitivity of land surface evapotranspiration to soil, vegetation and atmospheric variables has been carried out using the Biosphere Atmosphere Transfer Scheme (Gao et. al. 1996, Wilson et. al. 1987) and the Simple Biosphere Model (Sellers et. al. 1987). These examine in detail the sensitivity of evapotranspiration to atmospheric variables such as air temperature, wind speed, specific humidity and biophysical parameters such as the leaf area index and canopy resistance.

The theory for computing sensitivity of hydrological variables to land surface and atmospheric variables follows that of McCuen, (1973) and McCuen et. al, (1974). Beven (1979) was the first to use this theory and the Penman-Monteith equation for evapotranspiration along with observations to determine sensitivity with respect to vegetation type. Beven concluded that the differences in climate played a smaller role than differences in vegetation type on sensitivity of evapotranspiration. The study of Luxmoore et. al. (1981) included the effect of feedback in examining sensitivity of evapotranspiration to meteorological variables. They found that evapotranspiration depends less on net radiation than on dew point temperature and air temperature. These findings are opposite to that of Saxton (1975) which did not incorporate the effect of feedback and found that evapotranspiration depends mainly on net radiation. A more recent comprehensive study examining the sensitivity of evapotranspiration to land surface and atmospheric variables with and without feedback in the planetary boundary layer has been carried out by Jacobs et. al. (1992). Other studies include the role of correlated net radiation and relative humidity on errors in evapotranspiration (Ahn, 1996); the role of various soil, biophysical and atmospheric variables on the sensitivity of cumulative evapotranspiration at FIFE (First ISLSCP - International Satellite Land Surface Climatology Project, Field Experiment) and ABRACOS (Franks et. al. 1997a), and the use of bayesian estimates of uncertainty in fluxes (Franks et. al. 1997b). Ahn (1996) determined that since net radiation and relative humidity was correlated, the sensitivity of evapotranspiration to relative humidity was increased when the correlation between net radiation and relative humidity was taken into account. The work of Franks et. al. 1997a has shown that since many different sets of parameters may produce the same results the role of each parameter was evaluated using Monte-Carlo simulation thereby identifying redundant processes/parameters. The results of these simulations can then be used in conjunction with likelihood weights using the Bayes equation (Franks et. al. 1997b). This present work is different from the ones mentioned above as it examines the sensitivity of heat fluxes in the context of satellite data. Our interest in using satellite data is motivated by the fact that for computation of heat fluxes over large areas, one is limited by the number of ground observations. These sensitivity studies are crucial for climate studies (Mintz, 1984, Garratt, 1993, Shukla, 1982, Paturel et. al. 1995).

The subject of assimilation of soil moisture data or assimilation of meteorological data in order to estimate soil moisture more accurately is relatively a new area of study (McLaughlin, 1995). Recent advances in inverse methods (Entekhabi et. al. 1994, Lakshmi et. al. 1997c) have demonstrated the use of microwave satellite data in estimating soil moisture. The assimilation of soil moisture from low-level atmospheric variables using a mesoscale model (Bouttier et. al. 1993a,b) the assimilated soil moisture estimates help in the initialization of atmospheric models. Another class of methods use satellite estimates of surface temperature to adjust for the soil moisture (McNider et. al. 1994, Ottle et. al., 1994) and estimate with greater accuracy the surface fluxes and surface temperature.

Assimilation and sensitivity are related to each other. The sensitivity of computed hydrological variables (e.g. potential evapotranspiration) to air temperature and vapor pressure determines the amount of error that can occur in potential evapotranspiration due to errors in air temperature and/or vapor pressure. In assimilation, we are concerned in fixing the predicted values of soil moisture which depends on the computed potential evapotranspiration (which is in turn determined by the air temperature and vapor pressure dictated by the sensitivity).

In this paper we study the sensitivity of computed potential evapotranspiration to air temperature, vapor pressure, surface temperature, wind speed, the effect of errors in air temperature and vapor pressure on potential evapotranspiration. The results of the sensitivity study will help in hydrological modeling through the analysis of the effect of errors in meteorological variables on the water and energy balance. It will also help us in estimating measurement accuracy requirements/effects of meteorological variables on the accuracy of evapotranspiration. In addition we address the potential for using satellite measured surface skin temperature for updating model simulated soil moisture. The assimilated surface temperature updates the soil moisture, evapotranspiration, the other heat fluxes and the water balances of the land surface.

2 Theory

The land surface hydrology can be represented by a model having two layers as shown in Figure 1 (Mahrt and Pan, 1984; Lakshmi et. al., 1997a). The water balance for the model can be written as

$$\begin{aligned} z_1\frac{\partial\theta_1}{\partial t} &= P - E - R - q_{1,2} \\ z_2\frac{\partial\theta_2}{\partial t} &= q_{1,2} - q_{2,wt} - T \end{aligned} \tag{1}$$

where θ_1 and θ_2 are the volumetric soil moistures of layer 1 (with thickness z_1) and layer 2 (with thickness z_2), P is the precipitation, E is the bare soil evaporation, R is the surface runoff, T is the transpiration, $q_{1,2}$ is the moisture flow from layer 1 to layer 2 and $q_{2,wt}$ is the moisture flow from layer 2 to the water table. In this model, the bare soil evaporation is assumed to take place from the top layer only and the vegetation transpiration from the bottom layer only. The moisture flow from layer 1 to layer 2 ($q_{1,2}$) and the flow from layer 2 to the water table ($q_{2,wt}$) are modeled using the Richards equation accounting for the gravity advection and the moisture gradient. The bare soil evaporation and the vegetation transpiration are estimated using the supply and demand principle, i.e. if there is enough moisture to satisfy the potential value, the evaporation and transpiration occur at the potential rate, else they occur at a rate limited by the amount of available soil moisture. Potential evapotranspiration is defined as the evapotranspiration occuring in the absence of any restrictions in the supply of moisture or energy to the land surface. This is a important variable as the actual evapotranspiration is computed as being less than (in the case adequate amount of moisture or energy to satisfy the potential is unavailable) or equal to the potential evapotranspiration. In this paper we compute the potential evapotranspiration by solving the water and energy budgets of the land surface.

The energy balance equation for the land surface can be written as

$$R_{sd}(1-\alpha) + R_{ld} - \epsilon\sigma T_s^4 - \frac{\rho C_p}{\gamma(r_{av}+r_c)}(e_s(T_s) - e_a) - \frac{\rho C_p}{r_{ah}}(T_s - T_a) - \frac{\kappa}{D}(T_s - T_d) = 0 \quad (2)$$

Rewriting the above equation in more compact notation (replacing the resistance terms by an index),

$$R_{sd}(1-\alpha) + R_{ld} - \epsilon\sigma T_s^4 - LE_1(e_s(T_s) - e_a) - H_1(T_s - T_a) - G_1(T_s - T_d) = 0 \quad (3)$$

where R_{sd}, R_{ld} are the incoming shortwave and longwave radiation respectively. α and ϵ and σ are the albedo, emissivity and the Stefan-Boltzmann's constant respectively. ET is the latent heat of evapotranspiration; and T_s, T_a and T_d are the surface temperature, air temperature and the deep soil ($50cm$) temperature respectively. $e_s(T_s)$ and e_a are the saturated vapor pressure at surface temperature T_s and actual vapor pressure of the air respectively. ρ, C_p and γ are the density, specific heat and psychrometric constant of air; r_{av} and r_{ah} are the aerodynamic resistances to vapor and heat and r_c is the canopy resistance. κ and D are the thermal conductivity and the diurnal damping depth of the soil. The latent heat coefficient LE_1 can be viewed as a transfer coefficient (inverse of resistance) which translates the gradient in the vapor pressure $(e_s(T_s) - e_a)$ to latent heat flux. Similar interpretations can be made for H_1 and G_1. The aerodynamic resistances to vapor $(r_{av})$and heat (r_{ah}) are taken as equal to each other and are evaluated as (Brutsaert, 1982),

$$r_{av} = r_{ah} = \frac{1}{k^2 u_2}(\ln(\frac{z-d}{z_0}))^2 \quad (4)$$

where k is the Von Karman constant (0.4), u_2 is the $2m$ wind speed, z is $2.0m$ (the reference height), z_0 is the roughness length and d is the zero plane displacement. The canopy resistance is given by (Feyen et. al., 1980),

$$r_c = \frac{r_{min}^{st}}{\mathcal{L}} \quad (5)$$

r^{st}_{min} is the minimum stomatal resistance and $\mathcal{L}$ is the leaf area index. LE_1, H_1 and G_1 are variables that depend on surface resistance and thermal capacity of the ground respectively such that $LE_1(e_s(T_s) - e_a)$ equals the evapotranspiration flux LE, $H_1(T_s - T_a)$ equals the sensible heat flux H and $G_1(T_s - T_d)$ equals the ground heat flux. The latent heat coefficient LE_1 and the sensible heat coefficient H_1 are a function of the wind speed through the dependence of the aerodynamic resistances r_{av} and r_{ah} on wind speed (Figure 2). The heat storage is not included in Eqn. (2) as we have a thin upper layer (1.0cm) and a short time step (1 hour) in our computations. Therefore, the heat storage term is negligible. The latent heat coefficient is defined as the heat in Wm^{-2} per unit vapor pressure difference between the saturated surface vapor pressure and the ambient air vapor pressure in mb. The sensible heat coefficient is defined as the heat flux in Wm^{-2} per unit temperature difference between the surface and the air in K. The latent heat coefficient depends on the wind speed, roughness length (bare soil: z_0 =0.0001m, vegetation: z_0 = 0.07m), zero plane displacement (bare soil: d=0.0m, vegetation: d=0.25m), leaf area index and the minimum stomatal resistance (r^{st}_{min} = 100sm^{-1}). in addition to the factors listed above for vegetation cover. The solid line curves in the upper panel correspond to leaf area indices ranging from 1.0 to 8.0 at increments of 1.0. The dotted line corresponds to the bare soil. LE_1 increases linearly with wind speed in the case of bare soil (all the other factors being constant, u_2 is the only variable). In the case of soils with a vegetation cover, LE_1 increases linearly with wind speed for low wind speeds and then flattens out for higher wind speeds. This behavior is more apparent for lower leaf area indices ($\mathcal{L}$=1,2) than for the higher leaf area indices ($\mathcal{L}$=7,8). At higher $\mathcal{L}$ (such as $\mathcal{L}$=7, 8), LE_1 shows a very non-linear trend. The latent heat coefficient for vegetation covered soils LE_1 shows the greatest sensitivity to changes in wind speed for low values of wind speed and especially at larger values of leaf area index. At low values of wind speed and leaf area index, the latent heat flux (latent heat coefficient times the difference in vapor pressure), is small. As the wind speed increases (for small values of $\mathcal{L}$), there is not much change in LE_1. As the leaf area index increases, increase in wind speed helps increase the latent heat flux. Increase in wind velocity decreases the aerodynamic resistance and the increase in $\mathcal{L}$ decreases the canopy resistance simultaneously, thereby increasing LE_1. At low $\mathcal{L}$, the canopy resistance is a dominates over aerodynamic resistance, so increases in wind speed do not reduce the net resistance of the plant canopy to evapotranspiration. At large $\mathcal{L}$, the aerodynamic resistance dominates over the canopy resistance and increases in wind speed reduce the net resistance of the plant canopy to evapotranspiration. An interesting observation is that the bare soil LE_1 line intersects the vegetation cover LE_1 curves. It can be seen that the leaf area index at the point of intersection increases with increase in wind speed. The sensible heat coefficient (H_1) does not depend on the leaf area index and shows a linear trend for both the vegetation covered soil (shown by a solid line in the bottom panel of Figure 2) and bare soil (dotted line).

3 Sensitivity of Potential Evapotranspiration

In this section the sensitivity of potential evapotranspiration to air temperature, vapor pressure, surface temperature, wind speed and leaf area index will be studied. The inputs to the evapotranspiration calculation include air temperature, vapor pressure, wind speed, surface roughness and vegetation parameters. In this paper, we are concerned in ascertaining the effect of air temperature and vapor pressure on evapotranspiration. These two (air temperature and vapor pressure) are input to hydrological models using ground or satellite observations. Our emphasis is use of satellite based observations for air temperature and vapor pressure and the errors associated with it. The other inputs - wind speed, surface roughness and vegetation are equally important. However, these are not the focus of our study in this paper.

The sensitivity of the potential evapotranspiration to various land-atmosphere variables can be written as,

$$\Delta LE = \frac{\partial LE}{\partial T_s}\frac{dT_s}{dT_a}\delta T_a + \left(\frac{\partial LE}{\partial e_a} + \frac{\partial LE}{\partial T_s}\frac{dT_s}{de_a}\right)\delta e_a + \frac{\partial LE}{\partial \mathcal{L}}\delta\mathcal{L} + \frac{\partial LE}{\partial u_2}\delta u_2 \qquad (6)$$

The first two terms in the above expression are sensitivity of the potential evapotranspiration to surface air temperature and surface vapor pressure. The third and fourth terms show the sensitivity of the potential evapotranspiration to leaf area index and wind speed. In this paper, we investigate sensitivity of potential evaporation to surface and air temperature and vapor pressure only, therefore, terms 3 and 4 are not considered any further in our analysis. The first two terms can be further expanded using the dependence of saturation vapor pressure on temperature as (Raudkivi, 1979)

$$e_s(T_s) = 611exp\left(\frac{17.27T_s - 4714.7}{T_s - 35.7}\right) \tag{7}$$

where $e_s(T_s)$ is in Pa and T_s is in K. The slope of the saturation vapor curve with temperature is given by

$$e_s'(T_s) = \left(\frac{4098.16}{(T_s - 35.7)^2}\right) e_s(T_s) \tag{8}$$

Using the above equation and Eqn. 2, we can evaluate the first two terms (the sensitivity of potential evapotranspiration to air temperature and vapor pressure) of Eqn. 6 as

$$\begin{aligned} \frac{dLE}{dT_a} &= e_s'(T_s)LE_1\left(\frac{H_1}{4\epsilon\sigma T_s^3 + H_1 + G_1 + e_s'(T_s)LE_1}\right) \\ \frac{dLE}{de_a} &= -LE_1 + e_s'(T_s)LE_1\left(\frac{LE_1}{4\epsilon\sigma T_s^3 + H_1 + G_1 + e_s'(T_s)LE_1}\right) \end{aligned} \tag{9}$$

The sensitivity of potential evaporation to air temperature $\frac{dLE}{dT_a}$ and to vapor pressure $\frac{dLE}{de_a}$ is shown in Figure 3 as a function of wind velocity at 2m for bare soil and Figure 4 for vegetation covered soil with leaf area index of 4.0. In the figures, there are curves corresponding to surface temperatures of 273, 283, 293, 303, 313 and $323K$. The values of $\frac{dLE}{dT_a}$ and $\frac{dLE}{de_a}$ vary linearly with wind velocity at 2m for all surface temperatures for a bare soil. The maximum slope of $\frac{dLE}{dT_a}$ vs u_2 is for $323K$ and that for $\frac{dLE}{de_a}$ vs u_2 is for $273K$. The value of $\frac{dLE}{dT_a}$ increases with increase in surface temperature for a given wind velocity and the magnitude of $\frac{dLE}{de_a}$ decreases with surface temperature for a given wind speed. It can also be seen that the sign of the sensitivity of potential evapotranspiration is opposite between air temperature and vapor pressure, i.e $\frac{dLE}{dT_a}$ is greater than zero for all wind velocities and $\frac{dLE}{de_a}$ is less than zero for all wind velocities. The sensitivity factor $\frac{dLE}{dT_a}$ increases with temperature because $\frac{dLE}{dT_a}$ is directly proportional to $e_s'(T_s)$ which increases with temperature. This direct dependence outweighs the inverse dependence on temperature through the terms $4\epsilon\sigma T_s^3 + e_s'(T_s)LE_1$ in the denominator. The sensitivity with respect to vapor pressure, $\frac{dLE}{de_a}$ has a direct dependence on $e_s'(T_s)$ and an inverse dependence on $4\epsilon\sigma T_s^3 + e_s'(T_s)LE_1$ whose behavior is similar to $\frac{dLE}{dT_a}$, but the term $-LE_1$ dominates the sensitivity in Eqn. 9. Therefore, the increase in temperature increases the second term in Eqn. 9 and $\frac{dLE}{de_a}$ becomes less negative (hence decreases in magnitude). Increase in surface temperature (for a given wind velocity), increases the sensitivity of potential evapotranspiration to air temperature $\frac{dLE}{dT_a}$ and decreases the sensitivity of potential evapotranspiration to vapor pressure $\frac{dLE}{de_a}$. The sensitivity of potential evaporation to air temperature and vapor pressure for a vegetation covered soil of leaf area index equal to 1.0 and 8.0 for different values of wind velocity and surface temperature is shown in Table 1 and Table 2 respectively. These two tables show similar characteristics to the sensitivity curves in Figure 4 for L=4.0. There is increase in $\frac{dLE}{dT_a}$ with u_2 for a given T_s and an increase in $\frac{dLE}{dT_a}$ with T_s for a given u_2. The magnitude of $\frac{dLE}{de_a}$ increases with increase in u_2 for a given T_s and decreases with increase in T_s for a given u_2. In most of the offline (i.e. hydrological models run outside a Global Climate Model without a coupled atmospheric component and the atmospheric forcings are provided as input) land surface hydrology models, T_a and e_a are input variables which have some observation errors associated with them. We can examine the consequences of errors in the input air temperature and vapor pressure on the computed potential evapotranspiration. The sensitivities discussed here are based on hourly evapotranspiration. Typical numerical values of the sensitivities $\frac{dLE}{dT_a}$ and $\frac{dLE}{de_a}$ given at $293K$, $\mathcal{L}$=1.0, $4.0ms^{-1}$, are $0.024mmK^{-1}$ and -$0.018mmmb^{-1}$ respectively. This means that corresponding to an error in air temperature of $+1K$ (over any time period; in this case we use an hourly time step), the error in the computed potential evapotranspiration is $0.024mm$ and corresponding to an error in the vapor pressure of $+1mb$, the error in the potential evapotranspiration is -$0.018mm$. Therefore a consistent positive $3K$ error in hourly air temperature occuring for a period of 1 month would result in a cumulative error in potential evapotranspiration (computed hourly) of $5.18cm$. A consistent positive $5mb$ error in hourly vapor pressure for a period of 1 month would result in a $6.48cm$ cumulative error in potential evapotranspiration. If they both occur simultaneously, i.e a $3K$ error in hourly air temperature and a $5mb$ error in hourly vapor pressure (simultaneously), over a month, the resulting error in potential evapotranspiration would be 5.18-6.48=-$1.30cm$. It can also be seen from Tables 1 and 2, and Figures 3 and 4, the magnitude of $\frac{dLE}{dT_a}$ is greater than the magnitude of $\frac{dLE}{de_a}$ for $T_s \geq 293K$ for all wind velocities and vice-versa for $T_s < 293K$.

We will examine the errors in the potential evapotranspiration arising from errors in air temperature and vapor pressure inputs only, assuming the values of the wind speed, leaf area index, surface temperature and all radiation input are accurate. The signs of the two sensitivity factors $\frac{dLE}{dT_a}$ and $\frac{dLE}{de_a}$ are different as can be seen in Figures 3 and 4 and Tables 1 and 2. Therefore, there exists values of δT_a,δe_a pairs (δT_a^o,δe_a^o) that would result in zero ΔLE. This situation would also occur if the signs are same. In our case, T_a and e_a errors of the same sign tend to cancel, resulting in decreased sensitivity to vapor pressure errors to the extent that the errors are positively correlated. This relationship can be determined using Equation 6 and 9, i.e

$$\delta T_a^o = \frac{-\frac{dLE}{de_a}\delta e_a^o}{\frac{dLE}{dT_a}} \tag{10}$$

The solution of the above expression is presented in the form of lines of zero error in the computed potential evapotranspiration in Figure 5 and Figure 6 for a bare soil and leaf area index of 4.0 respectively for different surface temperatures (273, 283, 293, 303, 313 and $323K$) and $2m$ wind velocity u_2 (1, 2, 3, 4, 5, 8, 10, 15 and $20ms^{-1}$). It can be seen from these figures that the corresponding increase in δe_a^o with increase in δT_a^o is larger (slope of δe_a^o line) at higher values of the $2m$ wind velocity. At a given value of δT_a^o and u_2, the value of δe_a^o increases with increase in T_s. At $\delta T_a^o = 10K$, the range in δe_a^o caused by differences in wind velocity is the largest for the warmest surface temperature ($323K$). There are several differences between a bare soil and a vegetation covered soil as seen from Figures 5 and 6. The values of δe_a^o for a given u_2, δT_a^o and T_s is higher for a vegetation covered soil compared to a bare soil. This suggests larger errors in vapor pressure of the correct sign are needed over vegetation compared to bare soil so as to compensate for the air temperature errors, i.e. for the same error in air temperature, the error in computed potential evapotranspiration is larger over a vegetated soil than over bare soil. This can be seen on comparing the top panels (a) of Figures 3 and 4. The value of $\frac{dLE}{dT_a}$ for a given $2m$ wind velocity and surface temperature is larger for vegetation covered soil with $\mathcal{L} = 4.0$ as compared to bare soil (the vertical axis in the two figures have different ranges - 0-0.10mmK^{-1} in the case of the bare soil and 0-0.35mmK^{-1} in the case of the vegetation covered soil). In addition, for a given δT_a^o and T_s, the range in δe_a^o caused by differences in $2m$ wind velocity is much larger for the bare soil compared to the vegetation covered soil. This can be seen by comparing the bottom panels of Figures 3 and 4. The slope of $\frac{dLE}{de_a}$ vs u_2 for a bare soil is larger than for the vegetation covered soil. The values of δT_a^o, δe_a^o as a function of u_2 and T_s for leaf area index of 1.0 and 8.0, presented in Tables 3 and 4 respectively, show similar characteristics to that of leaf area index of 4.0.

4 Assimilation of Surface Temperature

4.1 Theory

Let T_s^m be the surface temperature computed by the land surface model and T_s^o be the observed surface temperature. Assuming equal validity of the two estimates, let the new assimilated temperature be $T_s^{'}$

$$T_s^{'} = \frac{T_s^m + T_s^o}{2} \tag{11}$$

The above expression is a simplistic representation. A more realistic relation would depend on a better representation of the model accuracy and the observation accuracy according to

$$T_s^{'} = \frac{(\Delta T_s^o)^2}{(\Delta T_s^m)^2 + (\Delta T_s^o)^2} T_s^m + \frac{(\Delta T_s^m)^2}{(\Delta T_s^m)^2 + (\Delta T_s^o)^2} T_s^o \tag{12}$$

where ΔT_s^o and ΔT_s^m are the errors in the observed surface temperature and the model derived surface temperature respectively assuming the errors are uncorrelated.

The assimilated surface temperature $T_s^{'}$ will have to satisfy the energy balance equation. Therefore, we can calculate the value of the evapotranspiration flux $ET^{'}$ that satisfies the same, i.e

$$ET^{'} = R_{sd}(1-\alpha) + R_{ld} - \epsilon\sigma T_s^{'4} - H_1(T_s^{'} - T_a) - G_1(T_s^{'} - T_d) \tag{13}$$

where $ET^{'}$ is the new evapotranspiration flux which is a result of the adjusted temperature $T_s^{'}$. $ET^{'}$ is a combination of the bare soil evaporation and the vegetation transpiration, (in depth units), we have

$$\frac{ET^{'}}{\rho_w L} = \frac{E^{'}}{\rho_w L} + \frac{T^{'}}{\rho_w L} \tag{14}$$

where $\rho_w$and L are the density and latent heat of vaporization for water. This new bare soil evaporation $E^{'}$ and vegetation transpiration $T^{'}$ are given by partition between where the bare soil evaporation is occuring and where the vegetation transpiration is occuring, i.e

$$E^{'} = ET^{'} \frac{W_1}{W_1 + W_2}$$

$$T' = ET' \frac{W_2}{W_1 + W_2} \tag{15}$$

W_1 and W_2 are the water holding capacities of layer 1 and layer 2 respectively along with the assumption that the bare soil evaporation E occurs from layer 1 only and the vegetation transpiration T occurs from layer 2 only (no roots in layer 1). The difference between the model computed and the new evapotranspiration flux ET' is given by

$$ET' - ET = \delta ET = -4\epsilon\sigma T_s^4 \delta T_s - H_1 \delta T_s - G_1 \delta T_s \tag{16}$$

where $\delta T_s = T_s' - T_s$, the difference between the assimilated surface temperature (Eqn. 11) and the model computed surface temperature (Eqn. 2). The partition of this difference in evapotranspiration δET into the difference for bare soil evaporation δE and the vegetation transpiration δT is given by,

$$\begin{aligned} \delta E &= \delta ET \frac{W_1}{W_1 + W_2} \\ \delta T &= \delta ET \frac{W_2}{W_1 + W_2} \end{aligned} \tag{17}$$

We have to change the soil moisture of layer 1 and layer 2 by $\delta\theta_1$ and $\delta\theta_2$ so that this new bare soil evaporation and vegetation transpiration hold good. In our present set up it is relatively easy as we have bare soil evaporation from the top layer only and transpiration from the bottom layer only. If,

$$\begin{aligned} \frac{\delta E}{\rho_w L} \frac{\Delta t}{\Delta z_1} &= \delta\theta_1 \\ \frac{\delta T}{\rho_w L} \frac{\Delta t}{\Delta z_2} &= \delta\theta_2 \end{aligned} \tag{18}$$

where Δt is the length of the time step in our land surface model and Δz_1 and Δz_2 are the thickness of layer 1 (from which the bare soil evaporation occurs) and layer 2 (from which transpiration occurs) respectively. We will adjust the layer 1 soil moisture and the layer 2 soil moisture by the correction factors $\delta\theta_1$ and $\delta\theta_2$ as

$$\begin{aligned} \theta_1' &= \theta_1 + \delta\theta_1 \\ \theta_2' &= \theta_2 + \delta\theta_2 \end{aligned} \tag{19}$$

θ_1' and θ_2' are the new soil moistures associated with the assimilated surface temperature T_s'

We have shown a novel way in which soil moistures can be updated using surface temperature. This method is completely general, it does not depend on the thickness of the soil layers or the type of physics used in the land surface model.

In case of a model with plant physiology and crop phenology, the above presented methodology can still be used. The equations will have to be re-derived as they are specific to the model framework.

4.2 Numerical results

In order to understand the numerical impact on changes in soil moisture using the above procedure, we have performed computations for a few scenarios. Corresponding to surface temperatures T_s of 273, 283, 293, 303, 313 and 323K, the energy balance factor $EB_f = 4\epsilon\sigma T_s^4 + H_1 + G_1$ is computed for a $2m$ wind speed of $4.0ms^{-1}$, zero plane displacement of $0.25m$ and roughness length of $0.07m$, thermal conductivity of $3.5Js^{-1}m^{-1}K^{-1}$ and diurnal damping depth of $0.5m$.

$$\begin{aligned}\delta\theta_1 &= (\delta T_s)(EB_f)\frac{1}{\Delta z_1}\frac{W_1}{W_1+W_2}\frac{\Delta t}{\rho_w L}\\ \delta\theta_2 &= (\delta T_s)(EB_f)\frac{1}{\Delta z_2}\frac{W_2}{W_1+W_2}\frac{\Delta t}{\rho_w L}\end{aligned} \tag{20}$$

We have chosen in our model $\Delta z_1 = 1.0cm$ and $\Delta z_2 = 99.0cm$. Using a residual soil moisture content θ_r of 0.02 and saturated soil moisture content θ_s of 0.50, W_1 = (0.50-0.02) 1.0 = $0.48cm$ and W_2 = (0.50-0.02) 99.0 = $47.52cm$ and the factors $\frac{1}{\Delta z_1}\frac{W_1}{W_1+W_2}$ and $\frac{1}{\Delta z_2}\frac{W_2}{W_1+W_2}$ are equal to $1.0m^{-1}$. This is a result of our choice of a hydrological model with a top thin layer of $1.0cm$ and a bottom layer of $99.0cm$. As a result of this simplification, the above expressions are identical for $\delta\theta_1$ and $\delta\theta_2$ as

$$\delta\theta_1 = \delta\theta_2 = (\delta T_s)(EB_f)\frac{\Delta t}{\rho_w L} \tag{21}$$

The values for ρ_w and L are $997kgm^{-3}$ and $2500KJkg^{-1}$ and Δt is 1 hour. Using the above expressions and the above values, the energy balance factor EB_f for the surface temperature range 273-323K is 85.44-$88.41Wm^{-2}K^{-1}$ and $\frac{\delta\theta_1}{\delta T_s}$ and $\frac{\delta\theta_2}{\delta T_s}$ equal to each other are in the range 1.23 X 10^{-4} - 1.28 X $10^{-4}K^{-1}$. This result shows that the impact of errors in surface skin temperature on the volumetric soil moisture content is very small. In the case of a $10K$ error in surface skin temperature, the error in the volumetric soil moisture contents $\delta\theta_1$ and $\delta\theta_2$ is 0.00125. This translates into a soil water depth of $0.00125cm$ for layer 1 and $0.124cm$ for layer 2. This is the error incurred in one hour. In case the difference between the model and the assimilated (satellite or ground) surface skin temperature is on the average $3K$ for each hour for 10 days, the error in volumetric soil moisture contents is 0.09. The corresponding error in the total soil water depths is $0.09cm$ in layer 1 and $8.9cm$ in layer 2. This could be as a result of the instrument error.

The impact of assimilation may appear small in the case of an hourly time step with this numerical example. However, on a cumulative basis, it adds up and results in a significant amount. The values of $\delta\theta_1$ are over a $1cm$ top layer. Changes in 0.05 volumetric soil moisture results in changes in infiltration, runoff and drainage flux from the top layer to the bottom layer. In addition, if the surface temperature changes with changes in evapotranspiration (which changes with soil moisture). The evolution of the boundary layer (the sensible heat flux depends on the partitioning of net radiation into sensible, latent and ground hear flux) depends on the surface temperature and sensible heat flux. Therefore, even small changes in the soil moisture would make drastic differences in the atmospheric circulation and boundary layer.

This paper offers a pilot study and a simple test to determine if the assimilation scheme has an impact on soil moisture computation. A more complete test using field observations of soil moisture will be carried out in the future.

5 Implications to use of Satellite Data

This section will utilize the sensitivity studies of potential evaporation to the input variables of air temperature and vapor pressure carried out in section 3. The cumulative value of potential evapotranspiration is an indicator of the amount of water loss from the soil column. This can be used in long term water balance studies to determine the cumulative runoff. Cumulative runoff is the difference between the cumulative precipitation and the cumulative potential evapotranspiration.

The importance of such a comparison is crucial because ground observations are absent/

inadequate in a major portion of the world. A comparison between the satellite and the ground observation derived potential evapotranspiration helps us to gain confidence in the trust we place in using satellite data alone for locations without ground observations. In this section we will compare the potential evapotranspiration estimates derived using ground observations with potential evapotranspiration estimates derived using satellite data.

A comparison of the colocated HIRS2/MSU (High Resolution InfraRed Sounder2/ Microwave Sounding Unit) derived air temperature and vapor pressure (Susskind et. al. 1997) with the corresponding observations over the region of the First ISLSCP (International Satellite Land Surface Climatology Project) Field Experiment (FIFE) (Lakshmi et. al. 1997b, 1997c) is presented in Table 5. The values are presented for the NOAA 10 (nominal equatorial nadir local observation time of 730 AM/PM) and NOAA 11 (nominal equatorial nadir local observation time of 130 AM/PM) satellites. The bias (satellite value minus in-situ ground value) averaged over the satellite observation period coincident with the FIFE data base (June 1987-November 1989) for the air temperature $\overline{\delta T_a}$ and vapor pressure $\overline{\delta e_a}$ is presented in Table 5. N_{obs} indicates the number of times over the two year period during which both the ground observations as well as the satellite retrieval were available. The ground observations are the average conditions over the entire FIFE site (FIFE follow-on) which is a $15km$X$15km$ region (Betts et. al., 1996). The FIFE conditions of surface temperature and the wind speed over the region were used in deriving the cumulative error in potential evapotranspiration $\Sigma\Delta LE$ using Equation 6. The $\Sigma\Delta LE$ are computed using the instantaneous values of the sensitivities of potential evapotranspiration to air temperature ($\frac{\partial LE}{\partial T_a}$) and vapor pressure ($\frac{\partial LE}{\partial e_a}$) and the instantaneous values of the errors in air temperature (δT_{ai}) and vapor pressure (δe_{ai}).

$$\Sigma\Delta LE = \Sigma\left(\left(\frac{\partial LE}{\partial T_a}\right)_i \delta T_{ai} + \left(\frac{\partial LE}{\partial e_a}\right)_i \delta e_{ai}\right) \tag{22}$$

Table 5 gives the various statistics associated with the computation of the cumulative potential evapotranspiration estimates. The mean error of air temperature $\overline{\delta T_a}$ and vapor pressure $\overline{\delta e_a}$ are presented for sake of reference purposes only as the cumulative value of the potential evapotranspiration error is computed by summation of the instantaneous errors in potential evapotranspiration which depend on the instantaneous errors in air temperature and vapor pressure.

We present the value of $\Sigma\Delta LE$ for three values of leaf area index, 0, 1, and 2 and the cumulative potential evapotranspiration ΣLE. The average leaf area index for FIFE is around 1.0 and the leaf area index varies between 0 (no vegetation) and 2. It can be seen that the cumulative error in computed potential evaporation ranges from $0.51mm$ for 129 observations for NOAA 11 AM in 1989 corresponding to bare soil to $20.76mm$ for 385 observations for NOAA 10 PM for 1987-1989 corresponding to $\mathcal{L}$=2.0. The value of $\Sigma\Delta LE$ increases with increase in $\mathcal{L}$ for all the cases of satellite overpasses. The sign of $\Sigma\Delta LE$ is negative for NOAA 10 AM and NOAA 11 PM. This is due to the negative bias in air temperature and a positive bias in vapor pressure. The sign of $\frac{\partial LE}{\partial T_a}$ is always positive and that of $\frac{\partial LE}{\partial e_a}$ is always negative (Figures. 3, 4 and Table 1,2), thereby giving rise to the negative $\Sigma\Delta LE$ for NOAA 10 and NOAA 11. It can also be seen that the magnitude of $\Sigma\Delta LE$ for NOAA 10 (AM and PM) is larger than that for NOAA 11 (AM and PM). This is due to the larger number (roughly two and a half times) the observations for the NOAA 10 cases versus the NOAA 11 cases. The yearly cumulative potential evapotranspiration error for $\mathcal{L}$=1.0 is -9.51, 14.36, 4.27 and $-24.81mm$ for NOAA 10 AM, PM, NOAA 11 AM and PM respectively. This shows that NOAA11 PM has the largest error per satellite observation. These errors must be interpreted in light of the cumulative potential evapotranspiration ΣLE calculated using the energy budget from the FIFE data set. This shows that the maximum potential evapotranspiration occurs during the NOAA11 PM overpass (which corresponds to the largest error per satellite observation), thereby making the percentage error very low (.645%). The largest percentage error is for the NOAA11 AM overpass (31.45%) which is coincident with the minimum value of potential evapotranspiration. The other two values of NOAA10 AM and PM fall inbetween with percentages of 2.6% and 16.2$ respectively. All of these error percentages are computed for the worst case scenario - $\mathcal{L}$, leaf area index corresponding to 2.0.

The results of this section have to be interpreted in light of particular applications, i.e is the cumulative error of a few mm in potential evapotranspiration significant? We believe that these errors are not significant and therefore warrant the use of satellite data in large scale hydrological modeling applications. The error in cumulative evapotranspiration of $20mm$ over a 385 day period translates to $0.05mm$ per day. The storage of the top layer of thickness $10mm$ and porosity of 0.5 is $5mm$. This means an error of 10% on a daily basis. Whereas such an error should be avoided, in absence of ground observations, satellite data will serve as a useful substitute.

6 Discussion and Conclusions

The present study deals with the effect of atmospheric and land surface variables on potential evapotranspiration. The differences between the bare soil and the vegetation case has been studied in detail. The effect of wind speed variations is greater on the bare soil than on vegetated areas. The latent heat coefficient for vegetation increases with increase in leaf area index and shows a greater sensitivity to wind speed at low values of wind speed. The latent heat coefficient for bare soil shows a linear increase with wind speed. The sensitivity of potential evapotranspiration to air temperature and vapor pressure has opposite signs for both bare soil and vegetation. In this work we have not carried out the sensitivity studies with respect to surface roughness. We recognize that surface roughness is an important input parameter in the estimation of aerodynamic resistance (Eqn. 4). Currently, surface roughness parameterizations are based on a fixed parameter list that relates the vegetation type to surface roughness length and zero plane displacement. However, with the advent of the Vegetation Canopy Lidar (VCL), (Dubayah et al. 1997), we should be able to get a better estimate of these roughness parameters. In addition, it would be possible to have the temporal variation of these parameters which change with the changes in the vegetation growth and decay.

We visualize that in the future, hydrological models will be driven mostly by satellite data input. Most of the regions of the world are data poor, i.e. lack ground observations needed for input into hydrological models. In order to carry out hydrological modeling of such areas, we would need spatially interpolated data (from nearby ground observations) or satellite data. Most of these stations are hundreds of kilometers away. Therefore, the results of spatial interpolation would be incorrect. Hence, satellite data provide us with a good alternative. The results of the sensitivity study have been used to ascertain the cumulative error in potential evapotranspiration based on using satellite data for air temperature and vapor pressure from NOAA10 and NOAA11 over the FIFE area in Manhattan, Kansas.

The most important conclusion of this study is the fact that the differences in instantaneous values of satellite derived air temperature and vapor pressure effect the final value of the cumulative potential evapotranspiration at a maximum of 0.1mm per satellite observation (Table 5, NOAA 11 PM, $\mathcal{L}$=2.0). This is indeed a very small error in potential evapotranspiration and therefore satellite data seems adequately suited for calculating this quantity. Therefore, satellite data can be used in hydrological studies on annual time scales without appreciable bias. The advantage to use of satellite data as opposed to ground observations are that they provide a much better spatial description of the variable in question. This is an important finding as most often, instantaneous comparisons of satellite data and ground observations result in large differences. In most of the cases, such comparisons (satellite vs ground data) are biased due to inadequate spatial sampling of the ground data and such comparisons are not warranted. In such situations, a process based comparison study like the one carried out for potential evapotranspiration in this paper may be a good alternative. In the absence of any errors in radiation and land surface characterization and with the assumption that the error in the computed potential evapotranspiration comes from errors in air temperature and vapor pressure only, there is a compensating effect of these errors on each other. This is used to determine the combination of air temperature and vapor pressure errors which result in zero error in computed evapotranspiration.

The assimilation of surface temperature shows a small effect on soil moisture on an hourly basis. However, carrying out an assimilation over a period of several days results in significant effect on soil moisture. Our method of surface temperature assimilation is quite similar to the methods of Ottle et. al. (1994) and McNider et. al. (1994) who have used surface temperature to adjust the model derived soil moisture. This method differs from the scheme of Bouttier et. al. (1993a,b) who use a regression between soil moisture and air temperature and relative humidity. They determine the optimal coefficients by minimizing the difference between the observations and the forecasts. The results of this paper can be used in the context of hydrological modeling to update soil moistures and verify the fact that these updated soil moistures are closer to the observations than the un-updated soil moistures.

Hydrological modeling has come a long way since the use of the bucket-model parameterizations in the late 1960s. Models today have most of the important physical processes parameterized and well represented. We now need to deal with the inadequacies of the monitoring and observing system to validate the accuracy of the model parameterizations. We need to have independent estimates of soil moisture through direct observations in order to ascertain if our updated soil moistures are indeed closer to the truth. However, observations of soil moisture are carried out on a routine basis in a very few number of locations and mostly in field experiments. There is a need for routine observations of soil moisture using satellite sensors. This will immensely help us in the field of land surface hydrological modeling.

References

[1] Ahn, H., Sensitivity of correlated input variables and propagated errors in evapotranspiration estimates from a humid region, *Water Resources Research*, 32 (8), 2507-2516, 1996

[2] Betts, A.K. and J.H.Ball, FIFE surface climate and site-average data set 1987-1989, Submitted to Journal of Atmospheric Science *submitted 3rd FIFE special issue*

[3] Beven, K., A sensitivity analysis of the Penman-Monteith actual evapotranspiration estimates, *Journal of Hydrology*, 44, 169-190, 1979

[4] Bouttier, F., J.-F.Mahfouf and J.Noilhan, Sequential assimilation of soil moisture from atmospheric low-level parameters. Part I: Sensitivity and calibration studies, *Journal of Climate*, 32, 1335-1351, 1993

[5] Bouttier, F., J.-F.Mahfouf and J.Noilhan, Sequential assimilation of soil moisture from atmospheric low-level parameters. Part II: Implementation in a mesoscale model, *Journal of Climate*, 32, 1352-1364

[6] Brutsaert, W., *Evaporation into the atmosphere, Theory, history and applications*, D.Reidel Publishing Co., 299pp, 1982

[7] Chen, F., K.Mitchell, J.Schaake, Y.Xue, H.L.Pan, V.Koren, Q.Y.Duan, M.Ek and A.Betts, Modeling of land surface evaporation by four schemes and comparison with FIFE observations, *Journal of Geophysical Research*, Vol. 101, No. D3, 7251-7268, 1996

[8] Dubayah, R. J.B. Blair, J.L. Bufton, D.B. Clark, J. Ja'Ja', R. Knox, S.B. Luthcke, S.Prince, J.Weishampel, The Vegetation Canopy Lidar Mission, proceedings of Land Information in the Next Decade II, Washington, D.C., published by *American Society of Photogrammetry*, Wash, D.C., CD-ROM, 1997.

[9] Entekhabi, D., H.Nakamura and E.G.Njoku, Solving the Inverse problem for soil moisture and temperature profiles by sequential assimilation of multifrequency remotely sensed observations, *IEEE Transactions on Geoscience and Remote Sensing*, Vol. 32, No. 2, 438-448, 1994

[10] Feyen, J.C. and D.Hillel, Comparison between measured and simulated plant water potential during soil water extraction by potted rye grass, *Soil Science*, 129, pp 180-185, 1980

[11] Franks, S.W., K.J.Beven, P.F.Quinn and I.R.Wright, On sensitivity of soil-vegetation-atmosphere transfer (SVAT) schemes: equifinality and problem of robust calibration, *Agricultural and Forest Meteorology*, 16, 63-75, 1997

[12] Franks, S.W. and K.J.Beven, Bayesian estimation of uncertainty in land surface-atmosphere flux predictions, *Journal of Geophysical Research*, Vol. 102, No. D20, 23991-23999, 1997

[13] Gao, X., S.Sorooshian and H.V.Gupta, Sensitivity analysis of the biosphere-atmosphere transfer scheme, *Journal of Geophysical Research*, Vol. 101, No. D3, 7279-7289, 1996

[14] Garratt, J.R., Sensitivity of climate simulations to land-surface and atmospheric boundary-layer treatments - A review, *Journal of Climate*, Vol. 6, 419-449, 1993

[15] Jacobs, C.M.J. and H.A.R.DeBruin, The sensitivity of regional transpiration to land surface characteristics - Significance of feedback, *Journal of Climate*, 5 (7), 683-698, 1992

[16] Lakshmi, V., E.F.Wood and B.J.Choudhury, A soil-canopy-atmosphere model for use in satellite microwave remote sensing, *Journal of Geophysical Research*, 102, D6, 6911-6927, 1997a

[17] Lakshmi, V., J.Susskind and B.Choudhury, Determination of Land Surface Skin Temperatures and Surface Air Temperature and Humidity from TOVS HIRS2/MSU Data, In press, *Advances in Space Research*, 1997b

[18] Lakshmi, V. and J.Susskind, Validation of TOVS land surface parameters using ground observations, *under preparation*, 1997c

[19] Lakshmi, V., E.F.Wood and B.J.Choudhury, Evaluation of SSM/I Satellite Data for Regional Soil Moisture Estimation over the Red River Basin, *Journal of Applied Meteorology*, Vol. 36, No. 10, pp 1309-1328, 1997d

[20] Luxmoore, R.J., J.L.Stolzy and J.T.Holdeman, Sensitivity of a soil-plant-atmosphere model to changes in air temperature, dew point temperature and solar radiation, *Agricultural Meteorology*, 23, 115-129, 1981

[21] Mahrt, L. and H.Pan, A two-layer model of soil hydrology, *Boundary-Layer Meteorology* 29, pp 1-20, 1984

[22] McCuen, R.H., The role of sensitivity analysis in hydrologic modeling, *Journal of Hydrology*, 16, 37-53, 1973

[23] McCuen, R.H., A sensitivity and error analysis of procedures used for estimating evaporation, *Water Resources Bulletin*, 10(3), 193-202, 1974

[24] McLaughlin, D., Recent developments in hydrologic data assimilation, *Reviews of Geophysics*, Supplement, 977-984, U.S. National Report to International Union of Geodesy and Geophysics 1991-1994, 1995

[25] McNider, R.T., A.J.Song, D.M.Casey, P.J.Wetzel, W.L.Crosson and R.M.Rabin, Towards a dynamic-thermodynamic assimilation of satellite surface temperature in numerical atmospheric models, *Monthly Weather Review*, 122, 2784-2803

[26] Mintz, Y., The sensitivity of numerically simulated climates to land-surface boundary conditions, 79-105, In *The Global Climate*, Edited by J.T.Houghton, Cambridge University Press, Cambridge, 1984

[27] Ottle, C. and D.Vijal-Madjar, Assimilation of soil moisture inferred from infrared remote sensing in a hydrological model over the HAPEX-MOBILHY region, *Journal of Hydrology*, 158, 241-264, 1994

[28] Paturel, J.E., E.Servat and A.Vassiliadis, Sensitivity of conceptual rainfall-runoff algorithms to errors in input data - case of the GR2M model, *Journal of Hydrology*, 168, 111-125, 1995

[29] Raudkivi, A.J., *Hydrology*, Pergamon Press, 1979

[30] Saxton, K.E., Sensitivity analyses of the combination evapotranspiration equation, *Agricultural Meteorology*, 15, 343-353, 1975

[31] Sellers, P.J. and J.L.Dorman, Testing the Simple Biosphere Model (SiB) using point micrometeorological and biophysical data, *Journal of Climate and Applied Meteorology*, Vol. 26, 622-651, 1987

[32] Shukla, J. and Y.Mintz, Influence of land-surface evapotranspiration on Earth's climate, *Science*, Vol. 215, No. 19, 1498-1501, 1982

[33] Susskind, J., P.Piraino, L.Rokke, L.Iredell and A.Mehta, Characteristics of the TOVS Pathfinder Path A Data Set, *Bulletin of the American Meteorological Society*, Vol. 78, No. 7, 1449-1472, 1997

[34] Wilson, M.F., A.Henderson-Sellers, R.E.Dickinson and P.J.Kennedy, Sensitivity of the Biosphere-Atmosphere Transfer Scheme (BATS) to the inclusion of variable soil characteristics, *Journal of Climate and Applied Meteorology*, 26, 341-362, 1987

List of Figures

u_2	T_s=273K		T_s=283K		T_s=293K		T_s=303K		T_s=313K		T_s=323K	
ms^{-1}	$\frac{dLE}{dT_a}$	$\frac{dLE}{de_a}$	$\frac{dLE}{dT_a}$	$\frac{dLE}{de_a}$	$\frac{dLE}{dT_a}$	$\frac{dLE}{de_a}$	$\frac{dLE}{dT_a}$	$\frac{dLE}{de_a}$	$\frac{dLE}{dT_a}$	$\frac{dLE}{de_a}$	$\frac{dLE}{dT_a}$	$\frac{dLE}{de_a}$
1	.005	-.013	.007	-.012	.011	-.010	.014	-.008	.017	-.006	.019	-.004
2	.007	-.017	.011	-.016	.017	-.014	.023	-.011	.030	-.009	.035	-.007
3	.008	-.020	.014	-.018	.021	-.016	.030	-.014	.039	-.011	.048	-.009
4	.009	-.021	.015	-.020	.024	-.018	.035	-.015	.046	-.013	.058	-.010
5	.009	-.022	.016	-.021	.026	-.019	.038	-.017	.052	-.014	.066	-.012
8	.010	-.023	.018	-.022	.029	-.021	.045	-.019	.064	-.017	.085	-.015
10	.010	-.024	.018	-.023	.031	-.022	.048	-.020	.069	-.018	.094	-.016
15	.011	-.025	.019	-.024	.033	-.023	.052	-.022	.078	-.020	.110	-.018
20	.011	-.025	.020	-.025	.034	-.024	.055	-.023	.083	-.022	.119	-.020

Table 1: Sensitivity of potential evapotranspiration to air temperature $\frac{dLE}{dT_a}$ (mmK^{-1}) and to vapor pressure $\frac{dLE}{de_a}$ $(mmmb^{-1})$ for different wind velocity and surface temperatures for $\mathcal{L}$ = 1.0

u_2	T_s=273K		T_s=283K		T_s=293K		T_s=303K		T_s=313K		T_s=323K	
	$\frac{dLE}{dT_a}$	$\frac{dLE}{de_a}$	$\frac{dLE}{dT_a}$	$\frac{dLE}{de_a}$	$\frac{dLE}{dT_a}$	$\frac{dLE}{de_a}$	$\frac{dLE}{dT_a}$	$\frac{dLE}{de_a}$	$\frac{dLE}{dT_a}$	$\frac{dLE}{de_a}$	$\frac{dLE}{dT_a}$	$\frac{dLE}{de_a}$
1	.008	-.023	.012	-.019	.016	-.014	.019	-.010	.021	-.007	.023	-.005
2	.016	-.041	.024	-.033	.031	-.025	.037	-.018	.042	-.013	.045	-.009
3	.023	-.055	.034	-.045	.045	-.034	.054	-.025	.062	-.018	.067	-.012
4	.028	-.068	.043	-.055	.057	-.043	.070	-.031	.081	-.022	.088	-.016
5	.033	-.078	.050	-.065	.069	-.050	.085	-.037	.099	-.027	.109	-.019
8	.044	-.102	.069	-.087	.097	-.070	.125	-.053	.148	-.039	.166	-.028
10	.049	-.113	.079	-.098	.113	-.080	.147	-.062	.177	-.047	.201	-.034
15	.059	-.134	.096	-.119	.143	-.101	.193	-.081	.240	-.063	.280	-.047
20	.065	-.147	.108	-.134	.165	-.116	.229	-.096	.292	-.076	.349	-.058

Table 2: Sensitivity of potential evapotranspiration to air temperature $\frac{dLE}{dT_a}$ (mmK^{-1}) and to vapor pressure $\frac{dLE}{de_a}$ $(mmmb^{-1})$ for different wind velocity and surface temperatures for $\mathcal{L}$ = 8.0

$\delta T_a^o(K)$	$u_2=1ms^{-1}$	$u_2=3ms^{-1}$	$u_2=5ms^{-1}$	$u_2=10ms^{-1}$	$u_2=15ms^{-1}$	$u_2=20ms^{-1}$	$T_s(K)$
1.0	.356	.410	.424	.433	.439	.440	273
3.0	1.068	1.231	1.272	1.298	1.317	1.320	
5.0	1.780	2.051	2.120	2.164	2.195	2.200	
8.0	2.848	3.282	3.392	3.462	3.512	3.520	
10.0	3.561	4.103	4.240	4.328	4.390	4.400	
1.0	.652	.756	.780	.801	.809	.813	283
3.0	1.957	2.267	2.341	2.403	2.427	2.439	
5.0	3.261	3.778	3.902	4.004	4.046	4.065	
8.0	5.217	6.044	6.244	6.407	6.473	6.504	
10.0	6.522	7.556	7.805	8.009	8.091	8.130	
1.0	1.115	1.312	1.360	1.406	1.418	1.427	293
3.0	3.344	3.937	4.079	4.219	4.254	4.280	
5.0	5.573	6.562	6.799	7.032	7.091	7.134	
8.0	8.917	10.500	10.878	11.251	11.345	11.414	
10.0	11.146	13.125	13.598	14.064	14.181	14.268	
1.0	1.829	2.191	2.281	2.360	2.382	2.397	303
3.0	5.487	6.574	6.844	7.079	7.145	7.192	
5.0	9.145	10.956	11.407	11.798	11.909	11.987	
8.0	14.632	17.529	18.251	18.877	19.055	19.179	
10.0	18.289	21.912	22.814	23.596	23.818	23.974	
1.0	2.864	3.514	3.669	3.798	3.828	3.856	313
3.0	8.593	10.541	11.007	11.393	11.485	11.569	
5.0	14.322	17.568	18.345	18.989	19.142	19.282	
8.0	22.915	28.108	29.352	30.383	30.627	30.852	
10.0	28.644	35.135	36.690	37.978	38.284	38.565	
1.0	4.386	5.398	5.675	5.894	5.957	6.020	323
3.0	13.159	16.193	17.026	17.681	17.870	18.061	
5.0	21.932	26.989	28.376	29.469	29.783	30.101	
8.0	35.091	43.182	45.402	47.150	47.652	48.162	
10.0	43.864	53.977	56.752	58.938	59.565	60.202	

Table 3: Vapor pressure errors (δe_a^o) in mb corresponding to air temperature errors (δT_a^o) for different $2m$ wind velocity and surface temperatures for a vegetated soil with leaf area

$\delta T_a^o(K)$	u_2=1ms^{-1}	u_2=3ms^{-1}	u_2=5ms^{-1}	u_2=10ms^{-1}	u_2=15ms^{-1}	u_2=20ms^{-1}	$T_s(K)$
1.0	.359	.412	.423	.434	.438	.439	273
3.0	1.077	1.235	1.269	1.302	1.313	1.318	
5.0	1.795	2.058	2.115	2.169	2.188	2.196	
8.0	2.872	3.292	3.385	3.471	3.501	3.514	
10.0	3.590	4.116	4.231	4.339	4.376	4.392	
1.0	.647	.756	.781	.801	.809	.812	283
3.0	1.941	2.268	2.344	2.404	2.426	2.436	
5.0	3.235	3.781	3.907	4.006	4.043	4.061	
8.0	5.176	6.049	6.251	6.410	6.469	6.497	
10.0	6.471	7.562	7.814	8.012	8.086	8.121	
1.0	1.113	1.314	1.366	1.405	1.420	1.426	293
3.0	3.340	3.941	4.097	4.214	4.259	4.277	
5.0	5.567	6.569	6.829	7.024	7.098	7.129	
8.0	8.908	10.510	10.926	11.238	11.357	11.406	
10.0	11.135	13.138	13.658	14.047	14.196	14.257	
1.0	1.833	2.194	2.281	2.356	2.382	2.394	303
3.0	5.500	6.581	6.842	7.067	7.147	7.182	
5.0	9.167	10.968	11.404	11.779	11.911	11.970	
8.0	14.667	17.548	18.246	18.846	19.058	19.152	
10.0	18.333	21.935	22.807	23.558	23.822	23.939	
1.0	2.877	3.511	3.656	3.792	3.840	3.864	313
3.0	8.630	10.534	10.967	11.377	11.521	11.592	
5.0	14.384	17.557	18.278	18.961	19.201	19.320	
8.0	23.014	28.091	29.244	30.338	30.722	30.911	
10.0	28.767	35.114	36.556	37.923	38.403	38.639	
1.0	4.346	5.419	5.661	5.883	5.964	6.002	323
3.0	13.038	16.258	16.984	17.649	17.891	18.005	
5.0	21.731	27.097	28.307	29.415	29.819	30.009	
8.0	34.769	43.355	45.292	47.064	47.711	48.014	
10.0	43.462	54.194	56.615	58.830	59.638	60.017	

Table 4: Vapor pressure errors (δe_a^o) in mb 29rresponding to air temperature errors (δT_a^o) for different $2m$ wind velocity and surface temperatures for a vegetated soil with leaf area

Satellite	AM/PM	$\overline{\delta T_a}(K)$	$\overline{\delta e_a}(mb)$	N_{obs}	$\Sigma LE(mm)$	$\Sigma\Delta LE(mm)$		
						$\mathcal{L}$=0	$\mathcal{L}$=1.0	$\mathcal{L}$=2.0
NOAA 10	AM	-0.74	0.71	301	428.6	-6.59	-7.85	-11.27
	PM	4.13	0.60	385	127.4	5.99	15.15	20.76
NOAA 11	AM	1.0	-0.33	129	6.45	0.51	1.51	2.03
	PM	-1.5	0.86	94	1557.6	-3.98	-6.39	-9.01

Table 5: Cumulative error in computed potential evapotranspiration for the First ISLSCP (International Satellite Land Surface Climatology Project) Field Experiment for different satellite overpasses and three different leaf area indices

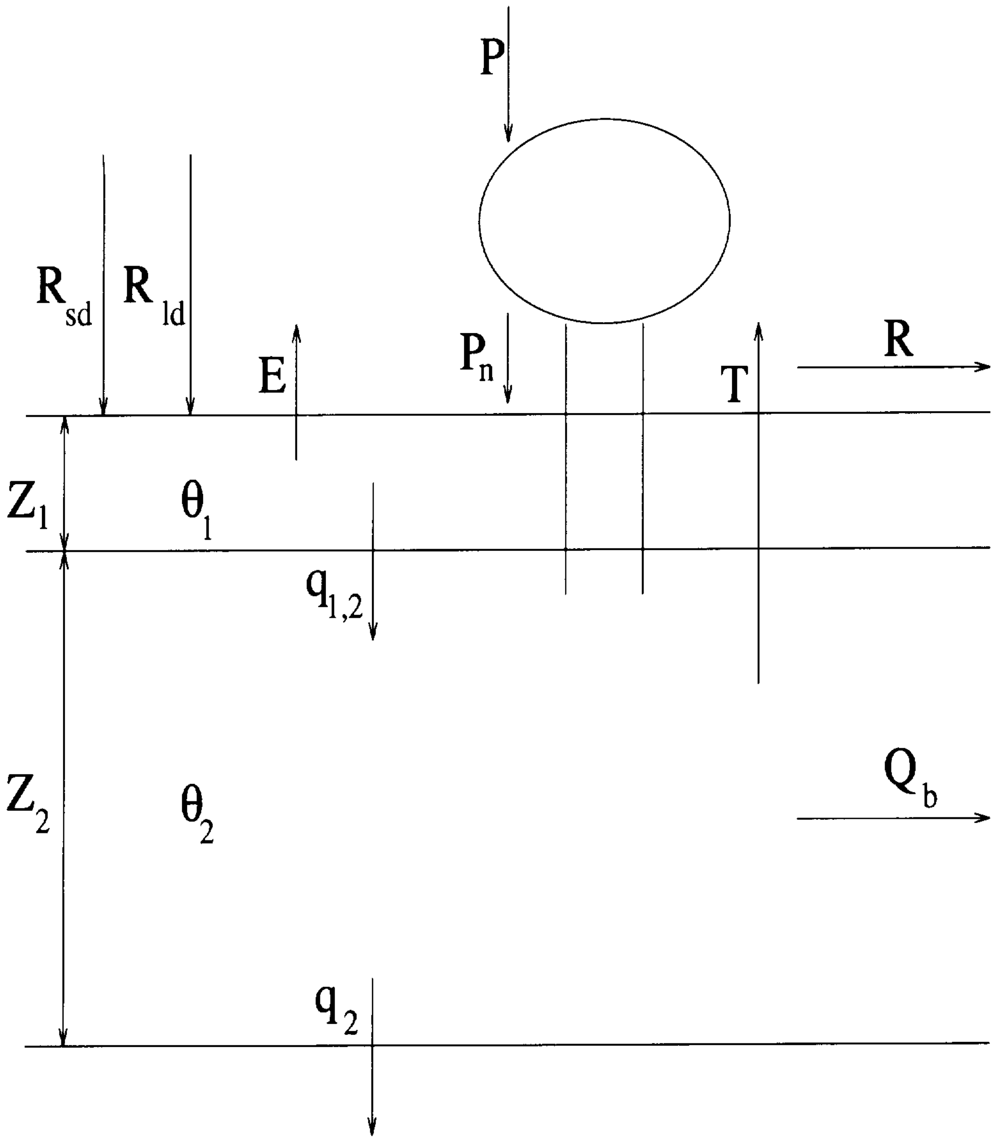
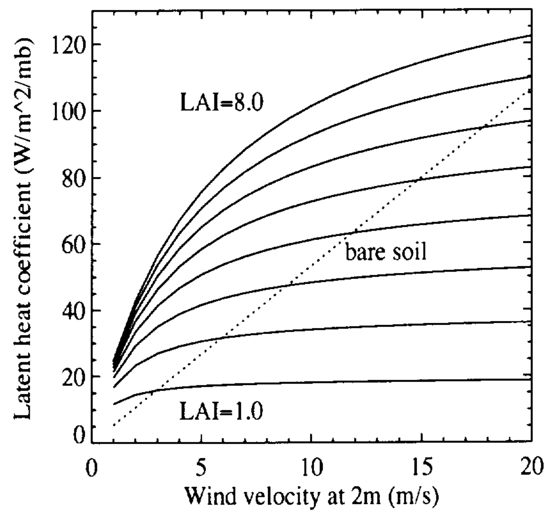


Figure 1: Water and energy balance in the land surface hydrology model



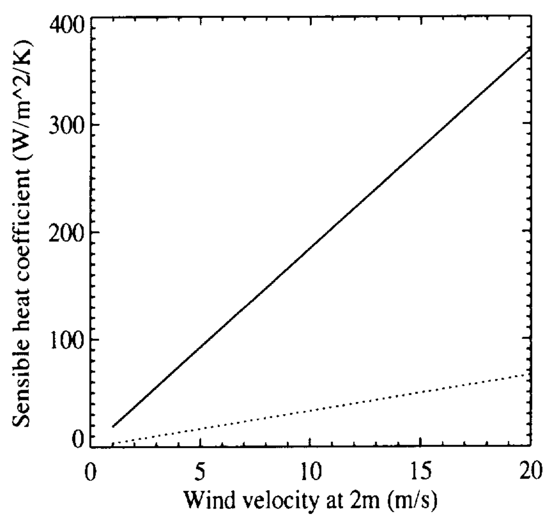


Figure 2: Variation of sensible and latent heat coefficients for vegetated and bare soil surfaces with wind speed

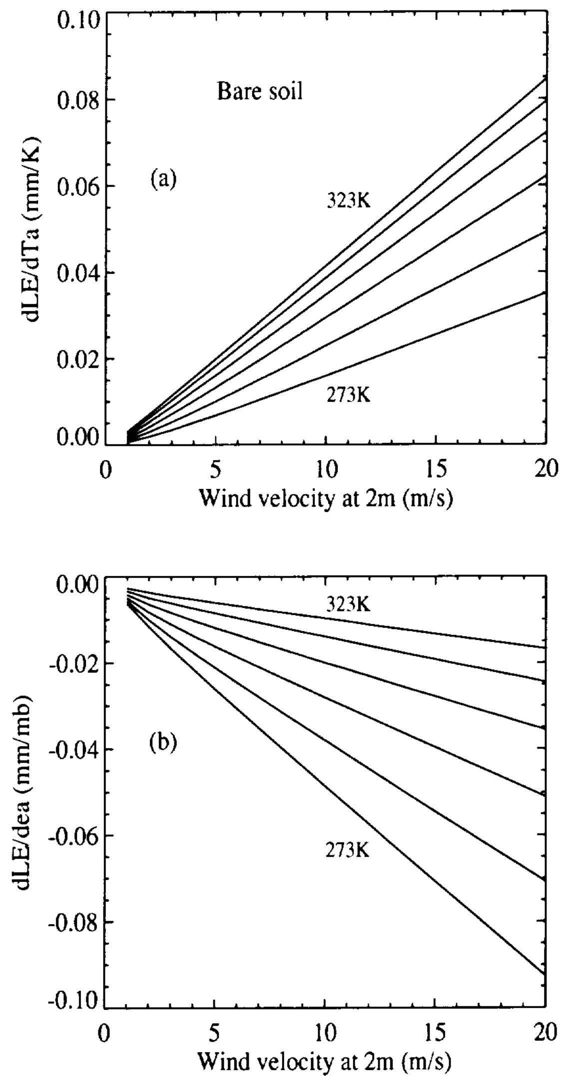


Figure 3: Sensitivity of bare soil potential evaporation to (a) Air temperature and (b) Vapor pressure as a function of wind speed for surface temperatures ranging from $273K$ to $323K$ at increments of $10K$