Annual Journal of Hydraulic Engineering, JSCE, VOL.44, 2000, February

ASSESSMENT OF ENERGY CONSIDERATION IN THE DESCRIPTION OF HYDROLOGICAL PROCESSES MODELING

Amila SILVA[1], Srikantha HERATH[2], Katumi MUSIAKE[3]

[1]Student Member of JSCE, M. Eng., Institute of Industrial Science, University of Tokyo
(4-6-1, Komaba, Meguro-ku, Tokyo 153-8505)
[2]Member of JSCE, Dr. of Eng., Guest Professor, Institute of Industrial Science, University of Tokyo
(4-6-1, Komaba, Meguro-ku, Tokyo 153-8505)
[3]Member of JSCE, Dr. of Eng., Professor, Institute of Industrial Science, University of Tokyo
(4-6-1, Komaba, Meguro-ku, Tokyo 153-8505)

Both land surface models and hydrological models use similar algorithms to describe water movement phenomena such as surface, sub-surface, ground water, etc, at a point. The fundamental difference between two schemes lies in evaporation estimation. Land surface models simulate actual evaporation using the energy balance whereas hydrologic models estimate evapotranspiration from the potential evaporation considering soil and plant properties such as soil moisture variation, leaf area index (LAI), plant root distribution, plant root depth, etc. This paper compares the performance of the land surface model developed at NCAR (LSM) and the hydrological model developed at IIS, University of Tokyo (IISDHM) against field observations from a hydro-meteorological observations (at 10 minutes interval) in an urban catchment. Both the LSM and IISDHM were able to simulate soil moisture variation adequately. Their long term evaporation estimates agree even though the short term patterns are different.

Key Words: *Land surface model, Hydrological model, Evapotranspiration, Energy consideration*

1. INTRODUCTION

Recent years have seen the development of numerous land surface models for global climate models. These models provide land-atmosphere exchanges of energy, moisture, and momentum, taking into account differences among vegetation and soil types in energy exchange processes. However, they differ greatly in the processes included in the model and how they parameterize similar processes.

Distributed physically based models give a detailed and potentially more correct description of the hydrologic processes in the catchment than do the other model types. Moreover, they are able to use as much as possible of the information and knowledge that is available concerning the catchment that is being modeled. The geographic information system (GIS) and remote sensing are two major techniques to support this modeling approach.

In this paper we compare the response of two hydrological process models namely Land surface model, Bonan et al., 1994[1] (herein referred as LSM), and Distributed hydrological model, Jha et al., 1997 [2] (herein referred as IISDHM) to examine the impact of energy consideration in different physical parameterizations. Both models use similar algorithms to describe water movement phenomena such as surface, sub-surface, ground water, etc. Estimation of evaporation is one of the fundamental differences between two schemes. Algorithm in IISDHM computes the evaporation in the sequence of intercepted water, transpiration from vegetation and evaporation from soil with the potential evaporation as the limiting value. On the other hand LSM considers the energy partition in addition to water availability.

2. MODEL DISCRIPTIONS

The major similarities and differences between

LSM and IISDHM are the following:

(a) Both models use numerical solution to the Richards' equation to calculate infiltration capacity.

(b) Both models calculate soil water flow for a six-layer soil but in addition LSM calculates soil heat fluxes.

(c) Both models calculate evaporation from soil and transpiration from vegetation but IISDHM calculates actual evaporation from potential evaporation considering soil and plant properties whereas LSM considers both energy and water balance. (**Fig. 1** and **Fig. 2** show the simple flow chart for two schemes)

(d) LSM computes all energy processes including latent heat, sensible heat and ground heat fluxes whereas IISDHM computes all hydrological processes including infiltration, ground water flow, over land flow and subsurface flow. These models are described in more detail as follows:

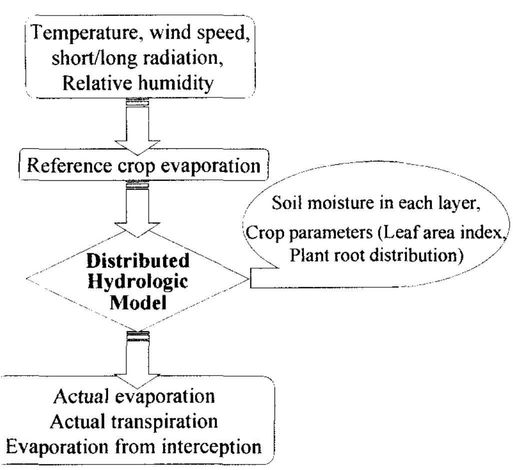

Fig. 1 Flow chart of IISDHM's evapotranspiration computation

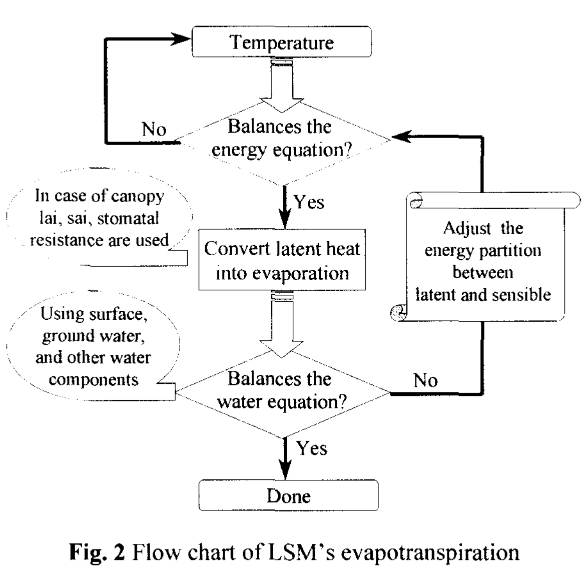

Fig. 2 Flow chart of LSM's evapotranspiration computation

(1) Land surface model (LSM)

LSM divides the land surface into a vegetation layer and the ground surface and solves simultaneously for the vegetation and ground temperatures that balance the surface energy budget.

In the simple case of non-vegetated surfaces, the net longwave radiative flux L (W m^{-2}), the sensible heat flux H (W m^{-2}), and the latent heat flux λE (W m^{-2}) depend on the ground temperature Tg. With the soil heat flux G (W m^{-2}) also a function of Tg, surface fluxes and temperatures are calculated by finding Tg that balances the energy budget.

$$-S_g + L(T_g) + H(T_g) + \lambda E(T_g) + G(T_g) + M = 0 \quad \textbf{(1)}$$

S_g is the net solar radiation (W m^{-2}) at the ground. L, H, and λE (positive towards the atmosphere) are by definition the ground surface fluxes. G (positive into the soil) is

$$G = \frac{2k_1}{\Delta z_1}\left(T_g - T_1\right) \quad \textbf{(2)}$$

Where k_l is the thermal conductivity (W m^{-1} K^{-1}), Δz_1 is the thickness (m), T_l is the temperature (K) of the first soil/lake/snow layer. M is the snow melt heat flux (W m^{-2}).

In the more complicated case of a vegetated surface, L, H, and λE are partitioned into vegetation and ground fluxes so that $L = L_v + L_g$, $H = H_v + H_g$, and $\lambda E = \lambda E_v + \lambda E_g$, where the subscripts "v" and "g" indicate vegetation and ground, respectively. These fluxes depend on vegetation T_v and ground T_g temperatures, which are determined as the temperatures that balance the canopy and ground energy budgets.

LSM solves simultaneously for the vegetation and ground temperatures that balance the surface energy budget. Surface fluxes are linearized about temperatures from the previous iteration to yield a linear system of two equations.

$$-S_v + L_v + H_v + \lambda E_v + \left(\frac{\partial L_v}{\partial T_v} + \frac{\partial H_v}{\partial T_v} + \frac{\partial \lambda E_v}{\partial T_v}\right)\Delta T_v = 0 \quad \textbf{(3)}$$

$$-S_g + L_g + H_g + \lambda E_g + \left(\frac{\partial L_g}{\partial T_g} + \frac{\partial H_g}{\partial T_g} + \frac{\partial \lambda E_g}{\partial T_g}\right)\Delta T_g = 0 \quad \textbf{(4)}$$

Where $\Delta T = T^{n+1} - T^n$ and superscript "n" indicates the iteration. Full heat process equations are given in Bonan et al., 1994[1].

Infiltration and Surface Runoff

The liquid water at the soil surface either infiltrates into the soil column q_{infl} (mm s^{-1}) or is lost as surface runoff q_{over} (mm s^{-1}). Surface runoff is;

$$R=\begin{cases} P+Q, & for \quad s\geq 1 \quad and \quad P>0 \\ P+Q-f^{*}, & for \quad s<1, \quad Q\geq f^{*}, \quad and \quad P>0 \\ P+Q-f^{*}, & for \quad s<1, \quad Q<f^{*}, \quad and \quad P>f^{*}-Q \end{cases} \quad \textbf{(5)}$$

Where P is throughfall (mm s^{-1}),

$Q = qmelt + qsdew$ (mm s^{-1}),

$s=\dfrac{\theta_1}{\theta_{sat}}$ is the water content of the first soil layer relative to saturation, and f^* is the infiltration (mm s^{-1}) which depends on s.

Infiltration is $I = P + Q - R$ **(6)**

The solution to R requires a physically realistic infiltration capacity that is easily integrated with respect to s.

$$f^{*}=k_{sat}vs+k_{sat}(1-v) \quad \textbf{(7)}$$

Where K_{sat} is the saturated hydraulic conductivity (mm s^{-1}),

$v=\dfrac{-(d\varphi/ds)}{0.5\Delta z_1}$ evaluated for s = 1, φ is the soil matrix potential (mm), and Δz_1=100 mm is the thickness of the first soil layer, with hydraulic properties defined at depth 0.5 Δz_1.

Soil Water

Soil water is calculated from the conservation equation

$$\frac{\Delta\theta\Delta z}{\Delta t}=-q_i+q_o-e \quad \textbf{(8)}$$

Where

θ is the volumetric soil water content (mm^3 mm^{-3})

Δz is the soil thickness (mm)

Δt is the time step (seconds)

e is the evapotranspiration loss (mm s^{-1})

q_i and q_o are the fluxes of water (mm s^{-1}) into and out of the soil (positive in the upwards direction).

Vertical water flow in an unsaturated porous media is described by Darcy's law as given below;

$$q=-k\left(\frac{\partial(\varphi+z)}{\partial z}\right)=-k\left(\frac{\partial\varphi}{\partial z}+1\right)=-k\left(\frac{\partial\theta}{\partial z}\frac{\partial\varphi}{\partial\theta}+1\right) \quad \textbf{(9)}$$

Where

k is the hydraulic conductivity (mm s^{-1}),

φ is the soil matrix potential (mm) and

z is the height (mm) above some datum in the soil column.

Setting e = 0, so that

$$\frac{\Delta\theta}{\Delta t}=-\left(\frac{q_i-q_o}{\Delta z}\right), \quad i.e. \quad \frac{\partial\theta}{\partial t}=-\frac{\partial q}{\partial z}, \quad \textbf{(10)}$$

Results in the Richards' equation

$$\frac{\partial\theta}{\partial t}=\frac{\partial}{\partial z}\left[k\left(\frac{\partial\theta}{\partial z}\frac{\partial\varphi}{\partial\theta}+1\right)\right] \quad \textbf{(11)}$$

This equation, with $e = q_{seva} + q_{tran}$ (soil evaporation and transpiration) and with the boundary conditions of q_{infl} as the flux of water into the soil and gravitational drainage $q_{drai} = k$ as the flux of water at the bottom of the soil column, is numerically implemented for a six-layer soil to calculate soil water.

Solution scheme for soil temperature is identical to soil moisture with only change of parameters (with soil thermal properties). Full description of land surface model is available at Bonan et al., 1994 [1)].

(2) Institute of industrial science distributed hydrologic model (IISDHM)

This model consists of four main components; namely interception, surface and river flow, subsurface flow and ground water flow. Spatial distribution of catchment parameters, rainfall input and hydrological response are represented in the horizontal plane by an orthogonal grid network and in the vertical plane by a column of horizontal layers at each grid. Some of the process which are essentially related with this study are described below:

Interception

The interception component calculates net rainfall reaching the ground through the canopy, the amount of water stored on the canopy and evaporation from the canopy. The canopy is considered to have a maximum surface storage capacity M_{ax}, which is filled by rainfall and emptied by evaporation and drainage. This capacity may be interpreted, as the minimum depth of water required to wet all canopy surfaces. I is depth of water on the canopy and calculated as.

$$I=C*LAI \quad \textbf{(12)}$$

Where C is Interception parameter (mm) and LAI is Leaf Area Index. The parameter C is dependent on vegetation. The evaporation from the canopy, *I_Ep* is calculated as;

$$I\_E_p = E_p \; if I \geq S_{\max} \; and$$
$$I\_E_p = E_p * I/S_{\max} \; if I < S_{\max} \quad \textbf{(13)}$$

Evapotranspiration

The actual evapotranspiration is calculated on the basis of potential evapotranspiration using actual soil moisture status in root zone, root distribution function and leaf area index (LAI) as other related parameters. The model adopted to calculate actual evaporation is the one described by Kristensen and Jensen et al.1971[3]. The potential evapotranspiration is divided into actual transpiration and actual evaporation. The actual transpiration takes place within the root zone of the crop, while the actual evaporation takes place only on the top layer of the soil. The actual transpiration E_{at} can be calculated as:

$$E_{at} = f(LAI) * f2(\theta) * RDF * E_p \quad \textbf{(14)}$$

Where, $f(LAI)$ is function of leaf area index; $f2(\theta)$ is function of moisture content at root depth level; *Ep* is potential evapotranspiration; *RDF* is function of root depth index.

The actual evaporation E_s is evaporation from a possible water reserve in the upper layer. Es is calculated as:

$$E_s = (E_p - E_{at}) f2(\theta) \quad \textbf{(15)}$$

Subsurface Flow

The subsurface flow component connects the surface flow and the ground water flow components. Soil moisture distribution in the unsaturated subsurface zone is calculated by solving three-dimensional Richard' equation. The *X* and *Y* components of the Richard' equation are solved explicitly for previous time step. Soil moisture flux due to *X* and *Y* components, extraction of moisture for transpiration and soil evaporation are introduced as sink terms at the node points in the root zone. Infiltration rates are determined by upper boundary, which might, be either flux controlled i.e. net rainfall or head controlled, in the case of ponding. The lowest node point included in the finite difference scheme depends on the phreatic surface level and allowance is made for the unsaturated zone to disappear in cases where the phreatic surface rises to the ground surface. The governing equation is written as,

$$C(\psi)\frac{\partial\psi}{\partial t} = \frac{\partial}{\partial z}[k(\psi)\frac{\partial\psi}{\partial z} + k(\psi)] + \frac{\partial}{\partial x}[k(\psi)\frac{\partial\psi}{\partial x}] + \frac{\partial}{\partial y}[k(\psi)\frac{\partial\psi}{\partial y}] - S \quad \textbf{(16)}$$

Where $C(\psi)$ is specific moisture capacity function; $K(\psi)$ is unsaturated hydraulic conductivity and S is source or sink term. Equation (26) is solved by implicit finite difference scheme. Full description of IISDHM are available at Jha et al., 1997 [2].

3. MODEL PERFORMANCES

(1) Study Area

Hydro meteorological observations (at "Futawa" station) in Ebi river basin, which is located in Chiba prefecture is selected for this study **(Fig.3)** Observations made at 10 minutes interval, from a micro-tower located at Futawa have been used for input data for model applications. Net radiation, air temperature, ground heat flux, relative humidity and wind speed data are used to compute potential evaporation using Kimberly Penman equation as described in Silva et al., (1999) [4]. Long wave radiation, relative humidity (to compute specific humidity) and precipitation, are used as input data for LSM. Soil moisture observations by TDR at three depths (0.1, 0.15, 0.25 and 1.0m) and soil temperature at two depths (0.03 and 0.05m) are used for comparison with the model results. The simulation has been carried out with this hydro-meteorological station data from 1st June to 10th September 1997 at 10 minutes interval.

(2) Soil moisture comparison

Fig. 4 shows the soil moisture fluctuation comparison between LSM and IISDHM with observations at the soil depth of 25cm. Results clearly show both models can capture the moisture movement with close agreement with the observations. However IISDHM soil moisture profile is more sensitive to rainfall inputs with high peaks matching with the observations. In general IISDHM results showed slightly higher soil moisture variations than that of LSM. The observed moisture profile showed a higher range of fluctuation than LSM, but in general lie in-between the two simulated values.

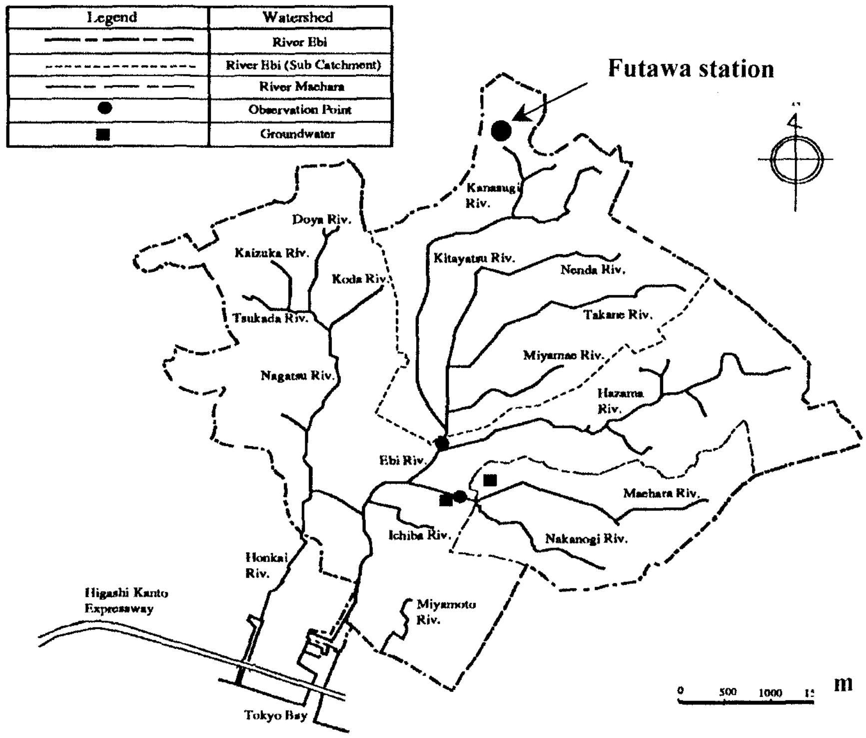

Fig. 3 Location of Futawa station at EbI river basin in Chiba prefecture

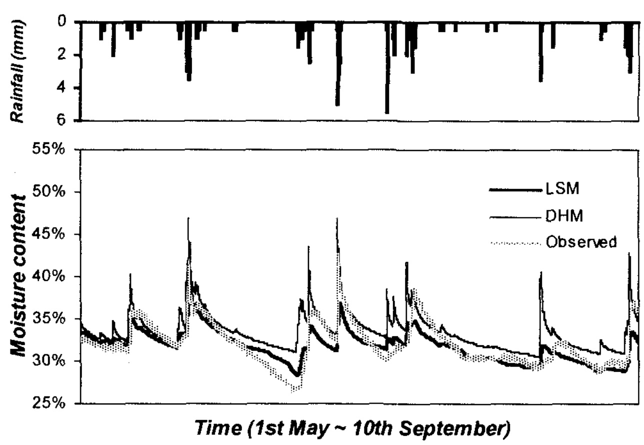

Fig. 4 Comparison of soil moisture fluctuation between LSM and DHM with observation at 25cm.

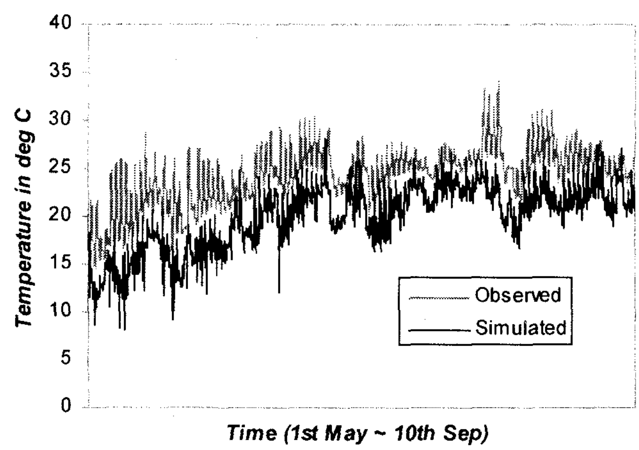

Fig. 5 Comparison of soil temperature fluctuation between LSM simulation and observation at 5cm.

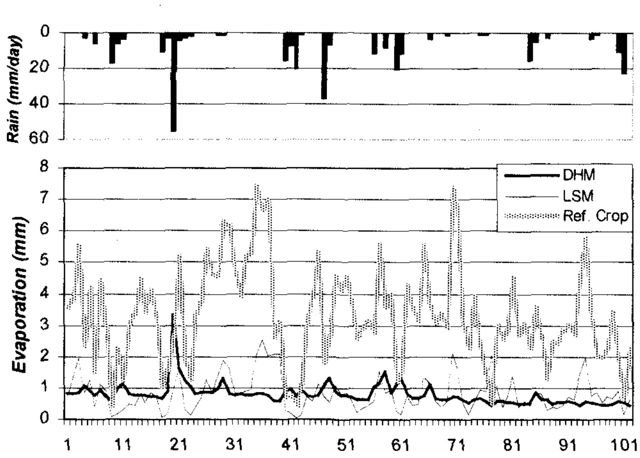

Fig. 6 Actual evaporation comparison for LSM and DHM with reference crop evaporation.

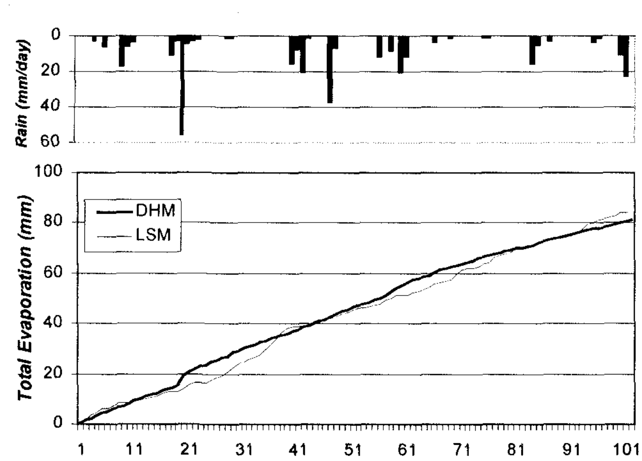

Fig. 7 Cumulative evaporation for LSM and DHM.

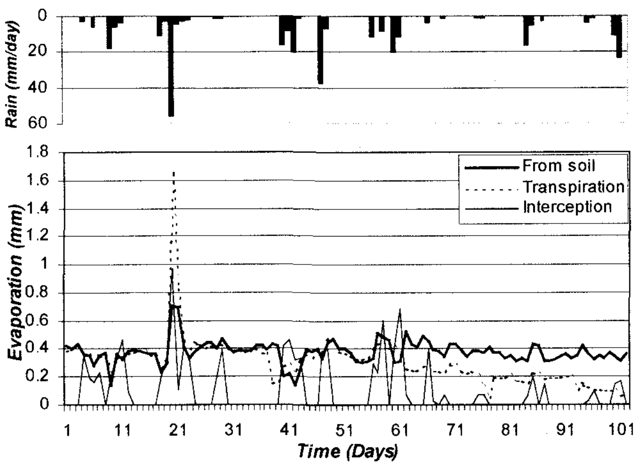

Fig. 8 Components of actual evaporation in IISHHM

(3) Soil temperature comparison

Fig. 5 shows the soil temperature comparison between LSM results and observed values at 5cm depth. The difference in values could be due to averaging through the layer depth (0cm ~ 10cm) in the case of LSM simulation and direct measurement at the 5cm depth in the case of observations. Though the values are exactly not matching with each other the actual (measured) temperature trend is closely followed by the LSM simulations.

(4) Actual evaporation comparison with reference crop evaporation

LSM and IISDHM actual evaporation in the daily scale are plotted against reference crop evaporation, which is used to compute IISDHM actual evaporation in the **Fig. 6**. With the rainfall input as a reference one can see IISDHM evaporation has high sensitivity to rainfall inputs. LSM daily evaporation and reference crop evaporation daily variation has good scaling relationship though reference crop evaporation is not used directly to estimate LSM evaporation. **Fig. 7** illustrates the cumulative daily evaporation comparison between models. However the cumulative evaporation is closely following each other irrespective of some peaks at different times. Energy and water based estimates (LSM) has more clear fluctuation similar to the reference crop evaporation (which is again estimated using energy and water based method). In contrast IISDHM results are more close to water availability (rain input).

(5) Components for actual evaporation simulated in IISDHM

Fig. 8 shows the daily distribution of different components of IISDHM for the entire simulated period. This graph clearly indicates the influence of LAI over the evaporation. Monthly LAI values are 3.5, 3.0, 1.5 and 0.7 for simulated months of June, July, August and September respectively. Transpiration reduces drastically with the reduction of LAI (after 61 days) whereas evaporation from soil remain more or less the same for entire simulated period.

4. CONCLUSIONS

As can be seen from the results above both Land surface model and Distributed hydrological model can simulate the actual moisture movement with good accuracy close to the observations. LSM simulations for soil temperature are promising as it captures the temperature variation trend adequately.

With the actual evaporation results, one can not say which methodology is closely related to the real world conditions. However it is understood that LSM is more sensitive to heat conditions and moisture from top most layer is taken away during dry periods. In comparison IISDHM is more sensitive to rainfall inputs. These results reveal that combination of both energy and water methodologies would describe the water movement phenomena between land and atmosphere more accurately.

Selecting the appropriate parameterizations such as LAI for both models is very important since simulation results depend on the similar physical description of surface and subsurface properties. These comparisons suggest that future development of models to describe hydrological cycle should pay more attention to the energy component of evapotranspiration process.

REFERENCES

1) Bonan, G.B. 1994. Comparison of two land surface process models using prescribed forcings. *Journal of Geophysical Research 99*:25803-25818.
2) Jha R., Herath S., Musiake K., Development of IIS Distributed Hydrological Model (IISDHM) and its application in Chao Phraya river basin, *41st Japanese conference on Hydraulics*, 1997.
3) Kristensen, K. and S. Jensen, A model for estimating actual evaporation from potential evapotranspiration. *Nordic Hydrology, 6*, pp. 70-88, 1975.
4) Silva A., Herath S, Hironaka S and Musiake K. Estimating Evaporation and its Spatial Distribution, *Proc. Annu. Conf. Japan Soc. Hydrol. And Water Resour.*, pp. 34-35, 1999.

(Received September 30, 1999)