Soil–Vegetation–Atmosphere Transfer Schemes and Large-Scale Hydrological Models (Proceedings of a symposium held during the Sixth IAHS Scientific Assembly at Maastricht, The Netherlands, July 2001). IAHS Publ. no. 270, 2001.

Coupled regional-scale hydrological–atmospheric model for the study of climate impact on Japan

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Abstract In order to be able to simulate long-term climate, it is necessary to model the evolution of both the atmospheric and hydrological variables in their fundamentally two-way interactive setting and to model the significant heterogeneity of land surface characteristics. A two-way feedback mode and the heterogeneity in a computational mesh are important keys to stable and realistic simulation. The coupling based on areally-averaged conservation equations was applied to a regional-scale atmospheric model for the main islands of Japan. Atmospheric–hydrological processes observed at Tsukuba station were compared with their counterparts simulated by the coupled model with a 20×20 km^2 resolution for the January 1989 historical period. The results of this comparison are quite satisfactory. Then, the model of Japan was run to simulate the climate change over Japan under the scenario of the doubling of CO_2 concentration in the atmosphere.

Key words Integrated Regional Scale Hydrologic/Atmospheric Model (IRSHAM); areally-averaged Green-Ampt model; soil water flow model; climate change

INTRODUCTION

Modelling a two-way interaction between the atmosphere and land surface hydrology and modelling the heterogeneity of land surface processes are key issues in atmospheric sciences and hydrology. Our modelling efforts started with a research project on climate change impacts on water resources over Japan. In order to perform long-term simulations of hydrological processes on a river-basin scale for climate change scenarios, a reliable and stably running mesoscale model was necessary, which we started to develop. After laboured trials, a model, named Integrated Regional Scale Hydrologic/Atmospheric Model (IRSHAM), was developed (PWRI, 1997). In the IRSHAM model, a two-way coupling scheme and areally-averaged conservation equations (Kavvas *et al.*, 1998) are employed. This made long-term realistic simulation possible on a regional scale.

This paper focuses, among several comprehensive model components of IRSHAM, on the land surface model with special attention to the soil moisture flow model based on the areally-averaged Green-Ampt equation to account for the heterogeneity of soils.

The core of the land surface moisture flow model is the spatially horizontally-averaged Green-Ampt model for infiltration and exfiltration over areally heterogeneous soils. This model is developed based on the study of Chen *et al.* (1994a,b). In the original study, only the infiltration with uniformly dry condition was considered, but IRSHAM has a newly developed set of spatially horizontally-averaged Green-Ampt equations, which can deal also with wet initial conditions and with exfiltration.

LAND SURFACE MODEL

The evapotranspiration rate (*ET*) can be estimated according to the moisture availability and the atmospheric conditions and by treating vegetation covered ground and bare soil differently. The total *ET* flux is then computed from a combination of the bare-ground evaporation (E_g), the direct evaporation from the canopy (E_{vd}), and the vegetation transpiration from the canopy (E_{tr}).

When the water availability parameter over bare soil, h, water availability parameters over vegetation, h_v and h_{vt} (Noilhan & Planton, 1989) and the bulk transfer coefficient, C_H (Stull, 1988) are available by the Monin-Obukhov surface similarity theory, the aerodynamic formulae can be used to estimate E_g, E_{vd} and E_{tr}. The aerodynamic formulae for evapotranspiration need for their computation the air density, the specific humidity of the first layer of the atmospheric model, the saturated specific humidity at ground surface temperature, and the bulk transfer coefficient (which is dependent upon the surface boundary layer stability of the atmosphere and is estimated by the Monin-Obukhov surface layer similarity theory). The water availability parameter for bare ground is related to surface water content, and is given by Noilhan & Planton (1989). However, for example, land surface temperature is determined from the heat budget of the land surface and is therefore dependent upon evaporation too. This is a part of the complex two-way nature of the land–atmosphere system, which should be solved in a coupled way (Kavvas *et al.*, 1998).

SOIL MOISTURE FLOW MODEL

The land surface model consists of three sub-models: the vegetation model for storage of water which is intercepted by vegetation; the evapotranspiration model; and the soil moisture flow model. Below we shall discuss the simple soil moisture flow component of the land surface model used in IRSHAM, which accounts for the heterogeneity of soils.

Usually, the depth of soil that interacts with the atmosphere is of the order of metres. Then, the scale of heterogeneity in the horizontal directions is much larger than the vertical scale of heterogeneity. Spatial variability of soil surface parameters governing moisture availability is very significant at the grid scales used in atmospheric models. Previous studies (Russo & Bresler, 1981a,b; Dagan & Bresler, 1983) have shown that the saturated hydraulic conductivity exhibits the maximum degree of spatial variability in many field soils. Then, one can treat the spatial variability of hydraulic conductivity as the primary cause of spatial variability in soil

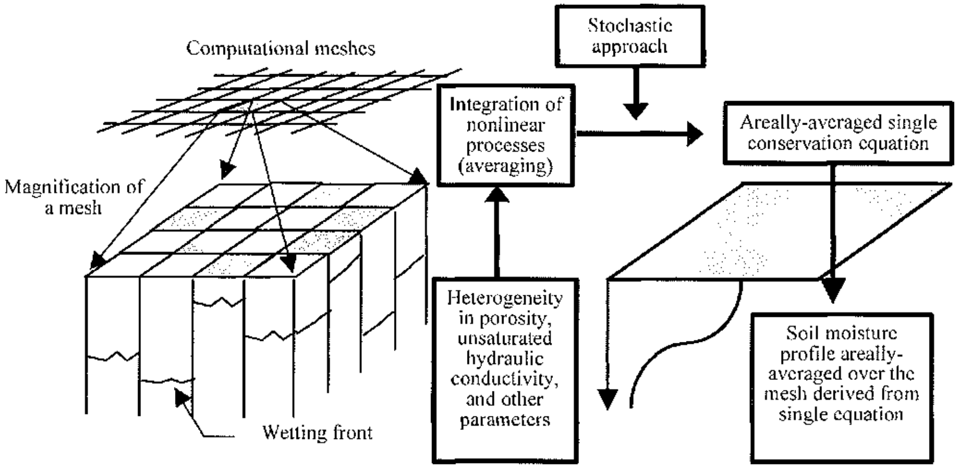

Fig. 1 Schematic description of averaging soil water profiles in a computational mesh.

water content profiles and treat the soil as composed of independent vertically homogeneous columns, which vary in the horizontal dimension. Under these conditions, Chen *et al.* (1994a,b) have developed the areally-averaged Green-Ampt models to delineate the areally-averaged water saturation profiles. Figure 1 shows the areal averaging concept. Because of significant spatial variability in soil parameters within a computational mesh, wetting front positions vary from point to point. It is clear that giving averaged parameters to a point location scale, the Green-Ampt equation or Richards' equation is unrealistic to describe the areally-averaged soil moisture profile. Instead, averaging the point-location-scale equation areally by stochastic approach yields a realistic profile. Based on the study by Chen *et al.* (1994a,b), a set of simulation schemes for infiltration and exfiltration under wet conditions was developed as land surface parameterization, which needs only the areally-averaged soil surface moisture content and areally-averaged water storage.

The vertical Darcian flux in terms of matric flux potential, i.e. $q_z = \dfrac{\partial \Phi}{\partial z} - K$, is used. In the Darcy flux representation, $K = K_s K_r$ and $\Phi = K_s \Phi_r$, K_s being the saturated hydraulic conductivity, K_r the relative hydraulic conductivity, and Φ_r the relative matric flux potential.

For the calculation of water flux at the land surface, the land surface is treated under primarily two boundary conditions: the infiltration condition and the exfiltration condition. As shown in Fig. 2, under infiltration the soil wetting front moves downward, increasing soil water saturation in the wet zone until the soil water saturation S reaches unity when the soil reaches saturation. Under the exfiltration condition the wetting front moves downward, losing soil water content in the wet zone.

The function $f_{Ks}(K_s)$, the probability density function or the frequency distribution function of saturated hydraulic conductivity K_s, is used to describe the spatial variability of the hydraulic conductivity within the soil of each model grid. A lognormal distribution of K_s is assumed in the IRSHAM model.

The water saturation S can be expressed by three different conditions as shown in Table 1. First, under the infiltration condition, such that water is applied to the

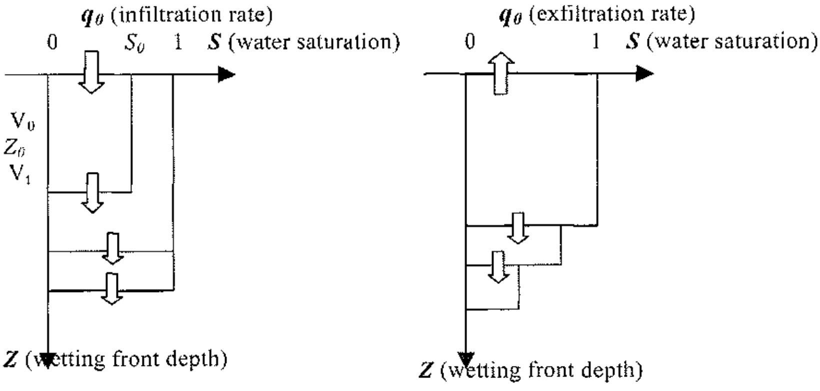

Fig. 2 Schematic description of soil moisture sub-model for land surface based on the Green-Ampt model.

unsaturated soil surface at a rate of q_0 ($q_0 > 0$), S can be expressed implicitly as a function of saturated conductivity, K_s, using the depth integrated Darcy's law and soil water continuity equations (Chen *et al.*, 1994a):

$$K_s = q_0 \frac{V_t}{V_t K_r(S) + S\Phi_r(S)} \tag{1}$$

where V_t is the normalized soil water storage without rainfall excess at time $t_0 + \mathrm{d}t$ under the permeable area, and

$$V_t = V_0 + \frac{q_0 \mathrm{d}t}{w_s - w_{\mathrm{wilt}}} \tag{2}$$

in which V_0 is the normalized soil water storage at time t_0, $\mathrm{d}t$ is the application time interval for q_0, and w_{wilt} is the soil water content at wilting point. Water saturation at time t_0 below depth Z_0 ($\equiv V_0/S_0$) is assumed to be zero where S_0 is the initial averaged water saturation at time t. For a fixed q_0 in infiltration condition and a fixed V_0, the soil surface with small saturated hydraulic conductivity will be saturated first. Therefore, soil surfaces can be grouped as saturated surface and unsaturated surface in terms of their saturated hydraulic conductivity values over the permeable area in each model grid.

Let K_{s0} be the maximum saturated hydraulic conductivity that the soil of saturated surface could have at time t_0, and K_{st} be the maximum saturated hydraulic conductivity that the soil of saturated surface could have at time $t_0 + \mathrm{d}t$. From equation (1), we obtain:

$$K_{s0} = q_0 V_0 / [V_0 + \Phi_r(1)] \text{ and } K_{st} = q_0 V_t / [V_t + \Phi_r(1)] \tag{3}$$

The top boundary condition of the moisture flow in soils with saturated surface is no longer the infiltration condition. It becomes the surface ponding condition. The soil moisture flow in soils with saturated surface at time $t_i + \mathrm{d}t$ is described by equations (4) and (5) in Table 1. While equation (1) is applicable to soils which are not saturated at time $t_0 + \mathrm{d}t$, equation (4) is for soils which have been saturated since time t_0 and equation (5) is for soils which were not saturated initially but are saturated at time

Table 1 Implicit equations for water saturation of a vertically homogeneous soil column.

Soil surface condition	Equation
Unsaturated	$K_s = q_0 \dfrac{V_t}{V_t K_r(S) + S\Phi_r(S)}$ (1) $V_t = V_0 + \dfrac{q_0 \cdot dt}{w_s - w_{\text{wilt}}}$ is the normalized soil water storage without rainfall excess at time t_0+dt under the permeable area, where V_0 is normalized soil water storage at time t_0, dt is the application time interval for q_0, and w_{wilt} is the soil water content at wilting point.
Saturated ($0 < K_s < K_{s0}$)	$K_s = \dfrac{\alpha}{dt\mu}\left[\dfrac{Z - V_0}{\alpha} - \ln\left(\dfrac{\alpha + Z}{\alpha + V_0}\right)\right]$ (4) where $\alpha = \dfrac{\Phi_r(S) - \Phi_r(S_i)}{K_r(S) - K_r(S_i)}$ and $\mu = \dfrac{K_r(S) - K_r(S_i)}{(w_s - w_{\text{wilt}})(S - S_i)}$ and S_i is initial water saturation
Transition from unsaturated to saturated during dt ($K_{s0} < K_s < K_{st}$)	$\dfrac{K_s}{q_0} = 1 - \exp\left[-\left(1 + \dfrac{V_t}{\alpha}\right)\dfrac{K_s}{q_0} + \dfrac{Z}{\alpha} - \ln\left(1 + \dfrac{Z}{\alpha}\right)\right]$ (5)

t_0 + dt. The areally-averaged surface saturation in the region of interest can be estimated from the three equations (1), (4) and (5), because the soil water content distribution in the region is completely described by these three equations when the saturated hydraulic conductivity distribution in the region is given.

In the unsaturated portion of the permeable area, the areally-averaged surface saturation is obtained by averaging S over all possible K_s values in the unsaturated area. In the saturated portion of the area, the saturation at surface is one.

The averaged surface saturation rate over the whole permeable area, including both saturated area and unsaturated area, is estimated as:

$$\overline{S} = A_s + \int_{K_{st}}^{\infty} S(K_s) f_{K_s}(K_s) dK_s \tag{6}$$

$$A_s = \int_0^{K_{st}} f_{K_s}(K_s) dK_s \tag{7}$$

where A_s is proportion of the saturated region in the permeable area, and $S(K_s)$ is the explicit function of K_s which is obtained by solving S from equation (1). If we view q_0 as the rainfall rate, the rainfall excess rate in the permeable area of each grid becomes

$$q_r = A_s\left(V_t - \overline{Z}_s\right)\frac{w_s - w_{\text{wilt}}}{dt} \tag{8}$$

where $\overline{Z}_s$ is the areally-averaged saturated depth in the saturated region within the permeable area of each model grid, and can be obtained from equations (3), (4) and (5).

Finally, the soil water storage per unit area at time t_0 + dt under the permeable area becomes:

$$W_r = \left(V_0 + \frac{q_0 - q_r}{w_s - w_{\text{wilt}}} \mathrm{d}t \right) \left[\frac{\bar{S} w_s + (1 - \bar{S}) w_{\text{wilt}}}{\bar{S}} \right] \tag{9}$$

Considering both permeable and impermeable areas, the rainfall excess per unit area in each model grid is:

$$R_{ex} = q_r \beta + q_0 (1 - \beta) \tag{10}$$

where β is the proportion of permeable area in each model grid.

The areally averaged soil surface water saturation and soil water storage under exfiltration conditions are obtained in a similar fashion as under the infiltration conditions.

APPLICATION TO JAPAN AND VERIFICATION

The land surface model including the soil moisture flow model was coupled with the boundary layer and 12-layer mesoscale atmospheric models in a two-way interaction, and applied to the main islands of Japan. The Japan model has a nested structure: the outer model has a 60×60 km^2 grid size and inner model has a 20×20 km^2 grid size. Historical simulation for the January 1989 period was carried out with this model. The US National Meteorological Center (NMC) dataset, the Comprehensive Ocean–Atmosphere Data Set (COADS), and the Japan AMeDAS dataset were used in the initialization for both the outer and the inner domain models. Daily precipitation, evaporation, net solar radiation, sensible heat flux, wind speed, dew point temperature, surface air temperature, and soil surface temperature, which were observed at Tsukuba station, were compared against their counterparts simulated by the coupled small domain model. Although the observed data are point-scale observations, dominant rainfall events are non-convective and have equal or larger spatial scale than the computational grid scale. The five rainfall events are generated on the right time intervals by IRSHAM as shown in Fig. 3. The simulated intensities have some discrepancy with the corresponding observations. This may be due to simulation error or to comparison of areally averaged rainfall simulation with point-scale observation. The order and daily variations of other processes are also simulated well, as shown in Fig. 3. Overall, the results are quite satisfactory for a model with no fitting effort.

CLIMATE CHANGE SIMULATION

The model of Japan was run to simulate the climate change over Japan under a scenario of the doubling of CO_2 concentration in the atmosphere. The boundary and initial conditions were interpolated from outputs of a Meteorological Research Institute GCM (MRI, 1984) for $1\times CO_2$ and $2\times CO_2$ scenarios. The climate change simulations show that there would be an almost uniform increase of 2°C throughout the main island, and that annual precipitation would decrease throughout the regions with the largest decrease occurring over the central mountain sector of Japan, as shown in Fig. 4.

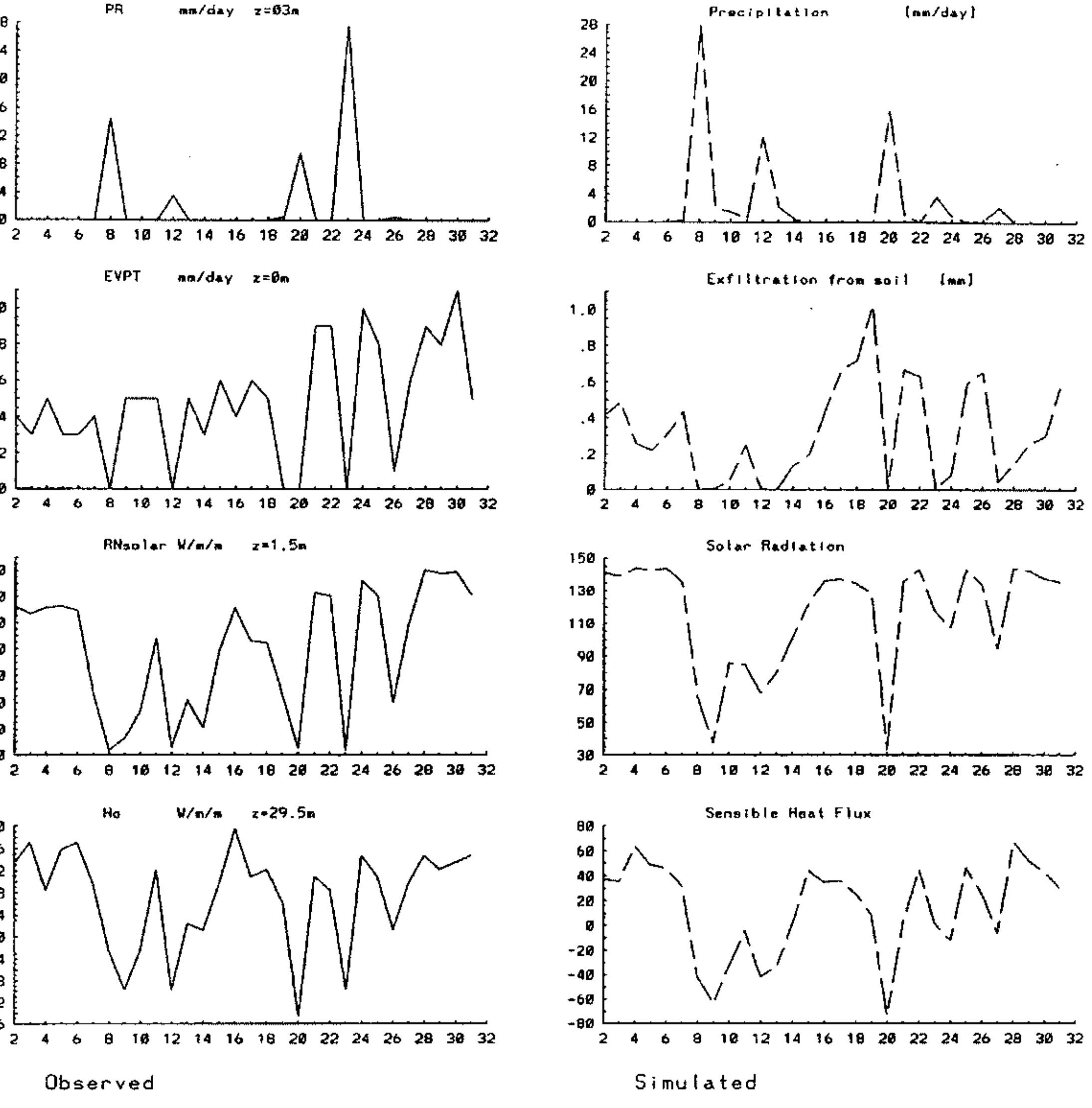

Fig. 3 Comparison of simulated and observed precipitation, evapotranspiration, solar radiation, and sensible heat flux in terms of daily time series at Tsukuba in January 1989.

CONCLUSIONS

A simple but physically-sound soil moisture flow model was developed and combined with the interception, *ET*, sensible heat flux, and radiation components of a land surface hydrology model. This land surface model was then integrated with boundary layer/atmospheric models in order to form a comprehensive Integrated Regional Scale Hydrologic/Atmospheric Model (IRSHAM). The soil moisture model is based on an areally-averaged Green-Ampt model and can account for the heterogeneity of

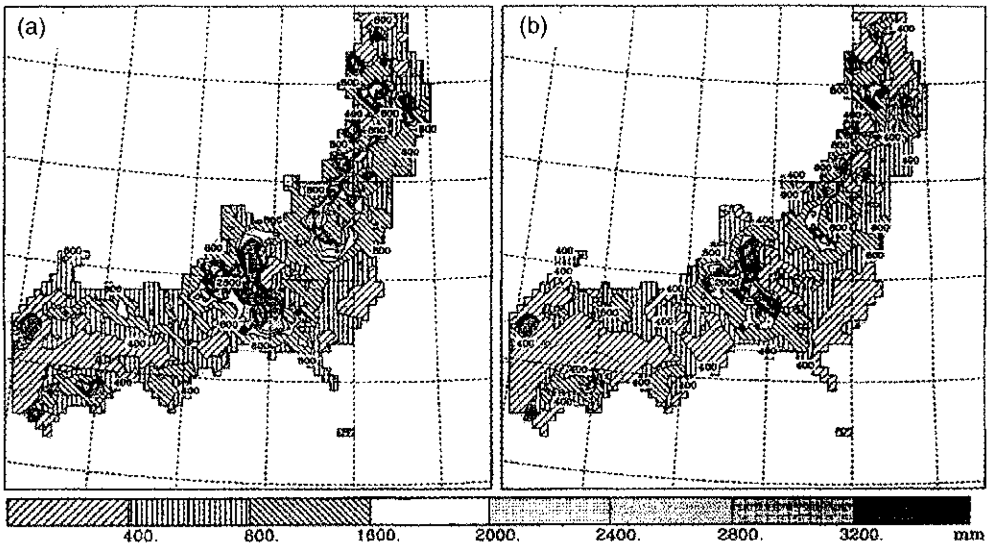

Fig. 4 Annual precipitation fields under (a) 1×CO_2 and (b) 2×CO_2 scenarios.

hydraulic parameters of soils within a computational mesh. Simulations for a historic period are compared with observations and the results are satisfactory.

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