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Impact of soil moisture movement schemes in a SVATS on the global climate of an AGCM

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Abstract Five numerical simulations with different versions of the schemes of soil moisture movement are conducted with the Meteorological Research Institute Global Spectral Model coupled with an improved Simplified Biosphere model to identify their effect on the model climate. One simulation with a new surface infiltration scheme improved the model's climate not only in the soil but also in the atmosphere, while the other experiments affected that in the soil only. The stronger persistence of the soil moisture anomaly at the third layer sustains a recycling system between evaporation and precipitation through soil water storage during a dry period, which improved the climate of the model. This study emphasizes the importance of the partition of precipitation into surface runoff and infiltration as well as that of the net radiation into latent heat and sensible heat at the Earth's surface.

Key words soil moisture movement; global climate; persistence; recycle; partition

INTRODUCTION

Many studies are dedicated to scale issues on horizontal heterogeneities at the grid scale of atmospheric general circulation models (AGCMs) (Nakaegawa *et al.*, 2000) because the scale of the land surface heterogeneity is much smaller than that of the AGCM grid. On the other hand, the vertical resolution of the soil in a soil–vegetation–atmosphere transfer scheme (SVATS) of an AGCM has rarely been discussed. The governing equation of the soil moisture movement is the Richards equation, which requires a fine vertical resolution because the matric potential changes drastically at the wetting front. A SVATS of an AGCM, however, usually has three layers and the thicknesses are typically 2 cm, some tens of centimetres, and one to two metres, respectively. Even if we use the most suitable physically-based equation, we cannot have realistic results without sufficient grid resolutions and an integration time step. This study examines the impacts of different versions of the scheme related to soil moisture movement on the global climate of an AGCM.

EXPERIMENT DESIGN

The AGCM used here is the Meteorological Research Institute (MRI) Global Spectral Model version 1 (GSM98) coupled with an improved Simplified Biosphere Model (L3SiB) (Numerical Prediction Division of the Japan Meteorological Agency, 1997; Ohizumi, 1997). The soil is divided into three layers: 2, 50–150 and 20–200 cm deep, respectively. This SVATS accounts for precipitation, snowpack formation and melting, water interception by and drip from the canopy, evapotranspiration, surface and underground runoff, and soil freezing and thawing. In the case of large vertical intervals, the representative relative unsaturated hydraulic conductivity differs significantly among different schemes. The conventional infiltration scheme produces oversaturation flow with rainfall of 10 mm h^{-1} when an integration time step is an hour, which contradicts the hydrological observations. The former depends on the vertical descretization, while the latter depends on the time step.

Five numerical experiments have been designed with different versions of the scheme, as shown in Table 1. The K1, R1 and D1 schemes are employed to alleviate negative influences of the insufficient vertical resolutions, while the I3 scheme is introduced to alleviate the defect in the conventional infiltration scheme mentioned above. The I3 scheme represents the implicit evaluation of the infiltration instead of the explicit one in the default scheme.

Five 3-year simulations were performed with MRI/GSM98 coupled with L3SiB. Monthly soil and atmospheric variables resulting from these simulations were used for this analysis. The model has a standard horizontal spectral resolution of T42 (approximately 2.8 × 2.8°) and 30 vertical levels to represent the atmosphere up to a height of about 0.5 hPa. Sea-surface temperature is prescribed with climatological monthly values.

Table 1 Description of the numerical experiments.

Experiment	Description
Control	Control default schemes are used for all processes.
K1	The larger saturated hydraulic conductivity of the two layers is given instead of the smaller one.
R1	The relative unsaturated hydraulic conductivity is computed from the arithmetic mean of the matric potential at the two layers instead of the flux potential method.
D1	The underground runoff is set as 1/10 of the default one.
I3	Rainfall can infiltrate into the second and third layers during rainfall events.

RESULTS

Figure 1 shows the combined atmosphere–land surface water balance of the control experiment and the differences from the control of each of its components (obtained by subtracting of the values of each experiment from those of the control experiment). All components of experiment K1 are almost same as those of the control experiment. The difference between the default and K1 schemes appears where the soil freezes, such as at high latitude and altitude, because soil characteristics are assumed to be uniform in depth. Such areas are limited and the influence on the global balance is insignificant. The R1 scheme reduces the moisture flux between the second and third layers, which makes their gradient increase. As a result, the underground runoff increases and reduces soil moisture at only the third layer. The D1 scheme directly reduces the underground runoff, but soil water storage does not increase much because it makes the surface layer wetter and the total amount of infiltration in experiment D1 smaller than that in the control experiment. Experiment I3 shows significant effects, not only

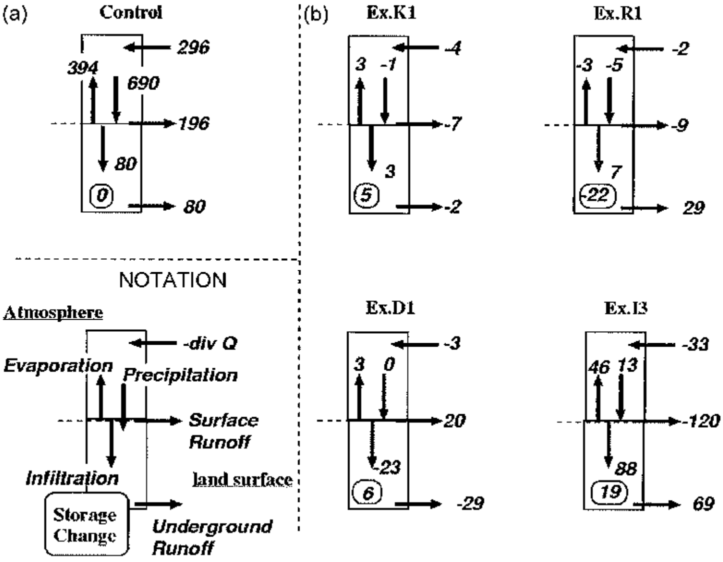

Fig. 1 Combined atmosphere–land water balance: annual mean values of (a) the control experiment model climate, and (b) the differences of each components compared to the control. Units are mm year^{-1} except water storage: mm/3 years.

on the soil moisture movement, but also on the atmospheric components such as precipitation and evaporation. The I3 scheme increases infiltration by a factor of two compared to the control experiment and soil water storage increases by 19 mm in three years. Precipitation increases by only 13 mm year^{-1}, while evaporation increases by 46 mm year^{-1}. As a result, water vapour convergence decreases 33 mm year^{-1} over the global terrestrial area, which corresponds to more than 10% of the control experiment. This suggests that the I3 scheme influences not only the surface hydrology but also atmospheric circulation at the global scale. In addition to the hydrological aspects, the higher bias of the ground temperature over the continents found in the control experiment is reduced. Here, these results are compared with estimates from the observation (Table 2). All components of experiment I3 are improved, while those of

Table 2 Terrestrial water balance of observational estimations and numerical experiments.

Variables	P	E	R	R_s	R_g	S_t	P_t
Control	690	394	276	196	80	-	1048
K1	689	397	267	189	78	5	1047
R1	685	391	296	187	109	22	1043
D1	690	397	267	216	51	23	1048
I3	703	440	225	76	149	19	1062
Sellers, 1965	720	410	310	-	-	-	1004
Korzun, 1978	800	485	303	-	-	-	1130
Oki & Musiake, 1995	704	467	237	-	-	-	915

P: precipitation (mm year^{-1}); E: evaporation (mm year^{-1}); R: total runoff (mm year^{-1});
R_s: surface runoff (mm year^{-1}); R_g: underground runoff (mm year^{-1}); S_t: difference in total soil water storage compared to control experiment at the end of simulation (mm/3 years);
P_t: global precipitation (mm year^{-1}).

the other experiments are not always improved. This indicates that the I3 scheme improves the model's climate more realistically.

DISCUSSION

The improvement of the model climate in experiment I3 is not necessarily related to the I3 scheme directly. Evaporation is controlled by both potential evaporation and soil water availability at the surface layer. However, potential evaporation is difficult to change between the control experiment and experiment I3, because the net radiation does not differ so much between them. A possible mechanism of increase of evaporation is the stronger persistence of soil moisture. During the non-rainfall periods, the Bowen ratio becomes larger due to the lower water availability. If the water is available during the dry periods, then the Bowen ratio can become small and, as a result, annual evaporation can increase.

The anomaly autocorrelation coefficients are a basic index used to estimate the magnitude of the persistence (Liu & Avissar, 1999). They are calculated as follows:

$$r_i = \frac{1}{N-\tau}\sum_{k=1}^{N-\tau}(W_{i,k} - \overline{W}_{i,k+\tau})(W_{i,k+\tau} - \overline{W}_{i,\tau})/\sigma^2 \qquad (1)$$

where τ is the lag time (in months), N is the length (months) of the simulation period (i.e. $N = 36$), $W_{i,k}$ is a time series of anomaly of soil moisture content at the ith layer for a sequential number of months k. The term $\overline{W}_{i\tau} = (1/(N-\tau))\sum_{k=1}^{N-\tau} W_{i,j}$ represents the average over $N-\tau$ months, and σ is the standard deviation.

Figure 2(a) shows the autocorrelation coefficients of soil moisture anomaly over the global terrestrial area at each layer in the control experiment (dashed lines) and experiment I3 (solid lines). The autocorrelation coefficients at the first and second layers in the control experiment are higher than those in experiment I3 during the first six months, while those at the third layer in the control experiment are lower than those

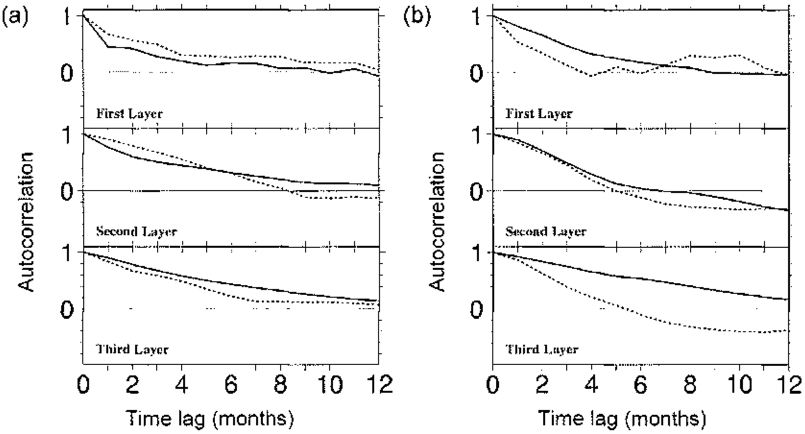

Fig. 2 Autocorrelation coefficients of control experiment and I3 experiment at each layer: (a) global and (b) Siberia. Dashed lines denote those of control experiment and solid lines denote those of experiment I3.

in experiment I3. Dry soil remains dry as long as it does not rain, while wetter soil reduces its water due to evaporation and infiltration. The soil at the first and second layers in control experiment is drier than that in experiment I3. This is because the anomaly autocorrelation coefficients of soil moisture in the control experiment are higher than those in experiment I3. Soil moisture at the third layer is not sensitive to the surface forcing such as evaporation and precipitation and, as a result, the larger the soil moisture, the stronger persistence the soil moisture has.

Therefore, the autocorrelation coefficients at the third layer in the control experiment are smaller than those in experiment I3. During dry periods, soil moisture at the third layer is moved upward and used effectively as evaporation, which can sustain higher evaporation rates even in a dry spell. This mechanism contributes to the increase of annual total precipitation in the middle of the continents and the lowering of the ground temperature. In contrast, the autocorrelation coefficients at all the layers in Siberia in experiment I3 are higher than those in the control experiment (Fig. 2(b)). Even wet soil does not lose water during winter in Siberia, since soil moisture cannot move due to freezing. Infiltration of snowmelt water in experiment I3 increases soil moisture because the I3 scheme increases infiltration directly. Potential evaporation is smaller in high latitude regions than in other regions because the available radiation is limited. Therefore the persistence in Siberia in experiment I3 is greater than that in the control experiment in all the layers. The mechanism in Siberia is a little different from that in the rest of the globe, but the I3 scheme enhances the persistence and improves the model's climate in Siberia as well as in other global terrestrial areas.

SUMMARY

Climate simulations produced with MRI/GSM98 coupled with L3SiB with five different schemes of soil moisture movement were conducted to identify their effect on the climate of the model.

Experiment I3 improves the model climate not only in the soil but also in the atmosphere, while the other experiments affect that in the soil only. The partition of the net radiation into latent heat and sensible heat is usually stressed as an important role of the land surface, but that of precipitation into the surface runoff and infiltration is also important because it influences the partition of the net radiation through soil water availability. The persistence of the soil moisture anomaly reveals a crucial key to this improvement from the viewpoint of the recycling system between evaporation and precipitation through soil water storage.

REFERENCES

Korzun, V. I. (1978) *World Water Balance and Water Resources of the Earth*, vol. 25 of Studies and Reports in Hydrology, UNESCO, Paris, France.

Liu, Y. & Avissar, R. (1999) A study of persistence in the land–atmosphere system using a general circulation model and observations. *J. Climate* **12**, 2139–2153.

Nakaegawa, T., Oki, T. & Musiake, K. (2000) The effects of heterogeneity within an area on areally averaged evaporation. *Hydrol. Processes* **14**, 465–480.

Numerical Prediction Division of the Japan Meteorological Agency (1997) Outline of the operational numerical weather prediction at the Japan Meteorological Agency. Appendix to *Progress Report on Numerical Weather Prediction*, 126. Japan Meteorological Agency.

Ohizumi (1997) Performance on a Simplified Biosphere Model with three soil layers implemented in MRI GSM, 75–90. *Tech. Report on Japanese Seasonal Prediction Skills, Meteorological Research Institute, Japan* (in Japanese).

Oki, T. & Musiake, K. (1995) Global atmospheric water balance and runoff from large river basin. In: *Scale Issues in Hydrological Modelling* (ed. by J. D. Kalma & M. Sivapalan), 411–434. John Wiley & Sons, Chichester, West Sussex, UK.

Sellers, W. D. (1965) *Physical Climatology*. University of Chicago Press, Chicago, Illinois, USA.