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Improvement of a Kinemative Wave-based Distributed Hydrologic Model to Predict Flow Regimes in Arid Areas

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Abstract Many studies have applied DHMs-KWSS concept (Distributed Hydrologic Models based on a Kinematic Wave approximation with Surface and Subsurface flow components) to predict river discharge for mountainous catchments in Japan. Some studies have showed that DHMs-KWSS represents well flow regimes of river basins with temperate climatic conditions, however it is difficult to explain flow regimes of arid basins. This study introduced the Hortonian overland flow caused by high rainfall in the dry season into a DHM-KWSS to explain the high discharge which happens at the beginning of rainy season in arid basins. The improved model represented high discharge in the beginning of rainy season better than the original DHM-KWSS structure.

Keywords *Distributed rainfall-runoff models, vertical infiltration, the Ayeyarwady river basin, arid basin, kinematic wave model*

Introduction

To develop a suitable hydrologic model to forecast river flows, dominant hydrologic processes in a catchment must be identified. In Japan, many river basins have steep slopes and are located in mountainous terrain, which causes rapid propagation of flood flows. These slopes are mainly covered by forests in which temperate climatic conditions keep soil moisture in the surface layer. Rainfall-runoff processes in these climatic conditions have been represented by some DHMs-KWSS (Distributed Hydrologic Models based on a Kinematic Wave approximation with Surface and Subsurface flow components) such as HydroBEAM (Kojiri *et al.* 2008) and OHDIS-KWMSS (Takasao *et al.* 1988, Sayama *et al.* 2006, Sayama and McDonnell 2009, and Kim *et al.* 2011).

Hunukumbura *et al*. (2012) used a DHM-KWSS to describe short-term flood events in several river basins located in multiple countries, encompassing a range of geographic and climatic conditions different from those in Japan. The results showed that the model was applicable to basins with steep slopes and a temperate climatic condition similar to those in Japan; however, it has difficulty in describing hydrologic behaviors in arid climates as well as those with mild topography. On the other hand, Tanaka and Tachikawa (2014) applied 1K-DHM (Tachikawa and Tanaka 2013) which is one of distributed hydrologic models with the DHM-KWSS structure to two river basins for long-term flow regimes. One has temperate climatic conditions as with river basins in Japan, and the other has different ones from river basins in Japan. It clarified that the DHM-KWSS model structure represents long-term flow regimes in a temperate climatic conditions and a large flood in arid basins; however, it is unable to explain long-term flow regimes in arid basins. The above two studies lead to the conclusion that the DHM-KWSS model structure is applicable to river basins with temperate climatic conditions and difficult to explain rainfall-runoff in arid basins.

Therefore, this study tries to improve 1K-DHM to better represent flow regimes in arid basins. In particular, sensitive responses of river discharge to rainfall in the dry season was targeted. The DHM-KWSS model structure provides all of rain water to subsurface flow because it is aimed at humid basins where humid surface soil layers have a large value of hydraulic conductivity. This characteristic does not match rainfall-runoff of arid basins, thus this study considered the Hortonian overland flow due to higher rainfall intensity than vertical infiltration capacity. The improved 1K-DHM was applied to the Ayeyarwady River basin located in Myanmar and reproduced the observed drastic change in river discharge in the beginning and middle of a rainy season.

Model description

1K-DHM is a distributed hydrologic model based

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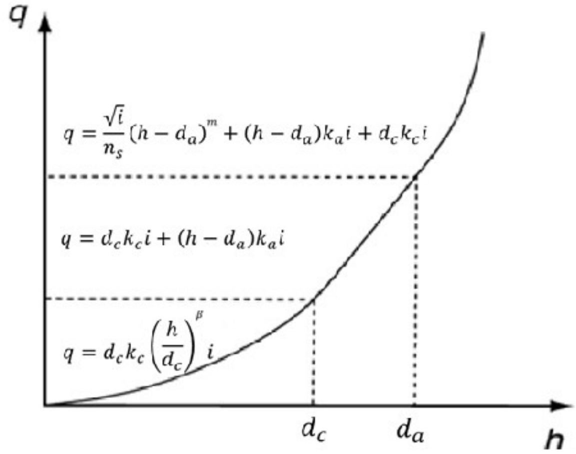

Fig. 1 Discharge-storage function of slope cells

on a kinematic wave flow approximation that considers subsurface flow. This model uses a kinematic wave model that describes surface and subsurface flow using a storage-discharge equation. Kinematic flow is applied to each cell within 1K-DHM, where flow is calculated using rainfall input to the cell and discharge from upper cells as the boundary condition. Flow direction is determined using topographical data provided by HydroSHEDS (http://hydrosheds.cr.usgs.gov).

Each cell has a slope component and a river channel component. Rain water is provided to a slope component, and runoff from a slope component is calculated by the following two equations, which consider both saturated and unsaturated subsurface flow components (Hunukumbura *et al*. 2012):

$$\frac{\partial h_{sub}}{\partial t} + \frac{\partial q_{sub}}{\partial x} = r - e \tag{1}$$

$$q_{sub} = \begin{cases} d_c k_c (h_{sub}/d_c)^\beta i & (0 \le h_{sub} \le d_c) \\ d_c k_c i + (h_{sub} - d_a) k_a i & (d_c \le h_{sub} \le d_a) \\ \sqrt{i}/n_s (h_{sub} - d_a)^m + (h_{sub} - d_a) k_a i + d_c k_c i & (d_a \le h_{sub}) \end{cases} \tag{2}$$

where h_{sub} is water stage; q_{sub} is discharge per unit slope width; r is rainfall intensity; e is intensity of evapotranspiration; d_c is an equivalent water stage to the maximum water content in the capillary pore; k_c is hydraulic conductivity when the capillary pore is saturated; β is an exponent parameter that describes the relationship between hydraulic conductivity and saturation; k_a is saturated hydraulic conductivity and $k_a = \beta k_c$ according to the continuity of the *q-h* relationship Equation (2); d_a is the water stage equivalent to the maximum water content in the effective porosity; and n_s is the Manning's roughness coefficient for saturation overland flow in a slope component. Equation (2) represents the q-h relationship for surface and subsurface soil layer as shown in Fig. 1. Runoff from a slope cell component is given to a river channel component as lateral inflow. River discharge is then calculated using the following kinematic wave equation:

$$\frac{\partial A}{\partial t} + \frac{\partial Q}{\partial x} = q_{sub} \tag{3}$$

$$Q = \alpha_c A^m \tag{4}$$

where A is the cross-sectional area; Q is discharge; r is rainfall intensity; e is evapotranspiration; and α_c and m are parameters. These parameters are determined as $\alpha_c = \sqrt{i}/n_c B^{-3/2}$ and $m = 3/5$, assuming a rectangular cross section of each cell. The parameters i, n_c, and B represent the slope gradient, Manning's roughness coefficient, and the width of a river channel. The value of B is determined by $B = bS^c$, where b = 1.06 and c = 0.69. These values were determined by considering the behavior of many observation stations across Japan.

Introduction of vertical infiltration into 1K-DHM

To consider the Hortonian overland flow, vertical infiltration component was introduced into a slope component of 1K-DHM. 1K-DHM-VI (1K-DHM that considers Vertical Infiltration) gives not all of effective rainfall $r - e$ but infiltration P to equation (1). P is defined as bellow:

$$P = \begin{cases} r - e & (r - e < f) \\ f & (r - e \ge f) \end{cases} \tag{5}$$

where, f is vertical infiltration capacity. Given larger effective rainfall $r - e$ than f, rain water that is unable to infiltrate $r - e - P$ contributes to the overland flow. f is defined as:

$$f = \begin{cases} k_c (h_{sub}/d_c)^\beta + k_0 & (h_{sub} < d_c) \\ k_c + k_0 & (h_{sub} \ge d_c) \end{cases} \tag{6}$$

As shown in equation (6), f represents vertical hydraulic conductivity that is function of water stage h_{sub}. To consider vertical infiltration through a crack on the extremely dry condition, minimal vertical

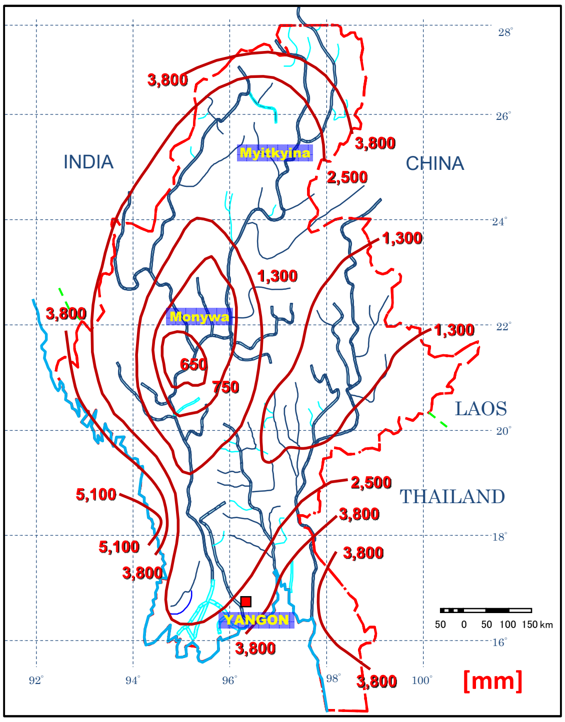

Fig. 2 Spatial distribution of a mean annual precipitation in the Ayeyarwady basin (The Kansai Electric 2008)

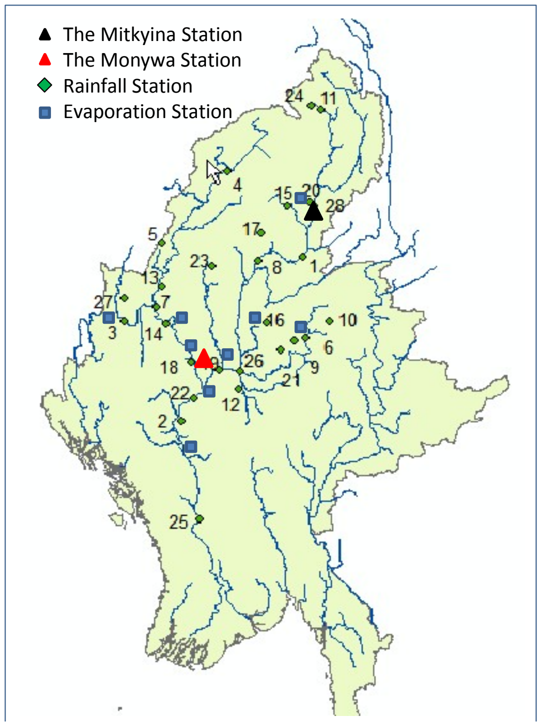

Fig. 3 water-shed boundary of the Ayeyarwady River basin and target stations (Myitkyina and Monywa)

hydraulic conductivity k_0 is added. Overland flow in a slope component is calculated by the following kinematic wave equation:

$$\frac{\partial h_{sfc}}{\partial t} + \frac{\partial q_{sfc}}{\partial x} = r - e - P \qquad (6)$$

$$q_{sfc} = \sqrt{i} / n_s h_{sfc}{}^m \qquad (7)$$

where h_{sfc} and q_{sfc} are water stage and discharge per unit slope width of the overland flow. q_{sub} and q_{sfc} contributes to a river channel component as lateral inflow. Consequently, the continuity equation of a slope component (1) and that of a river channel component (3) are changed as bellow:

$$\frac{\partial h_{sub}}{\partial t} + \frac{\partial q_{sub}}{\partial x} = P \qquad (8)$$

$$\frac{\partial A}{\partial t} + \frac{\partial Q}{\partial x} = q_{sub} + q_{sfc} \qquad (9)$$

Applying 1K-DHM-VI to the Ayeyarwady River basin

Climatic and topographic conditions

The Ayeyarwady River basin is characterized by drastic spatio-temporal variability of climatic conditions from north to south of the basin. The spatial distribution of a mean annual precipitation in Myanmar is shown in Fig. 2. Fig. 2 shows that a mean annual precipitation largely varies from north to south. Affected by topography, a mean annual precipitation in the southern and the western part of the basin amounts more than 3000 to 5000 mm and that in the central area is less than 1000 mm. The whole area of the basin has clear dry and rainy seasons. The rainy season starts from middle of May through October, and the dry season covers November through early May.

This study simulated flow regimes at the Myitkyina and Monywa stations in the Ayeyarwady River basin. The location of the two stations and water-shed boundary of the Ayeyarwady River basin are shown in Fig. 3. Comparing Fig. 2 and Fig. 3, we found that an upstream area of the Myitkyina station has much rainfall, while the amount of annual precipitation in an upstream of the Monywa station is smallest in the whole basin.

Calibration framework

The parameters of 1K-DHM were calibrated to reproduce flow regimes at both the stations in 2001 by using the SCE-UA algorithm (Duan *et al*. 1994). Daily precipitation at 28 stations and potential

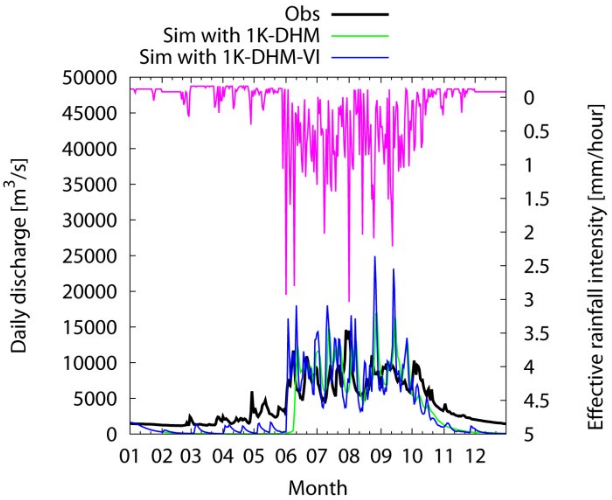

Fig. 4 Hydrograph of observed (thick line) and simulated discharge in 2001 at the Myitkyina station (effective rainfall means rainfall minus evapotranspiration)

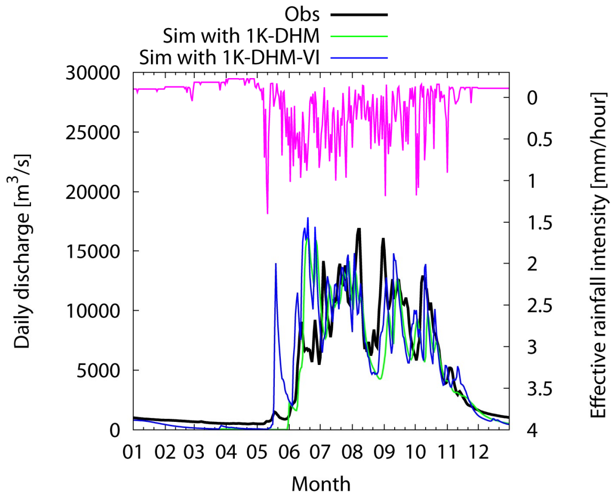

Fig. 5 Hydrograph of observed (thick line) and simulated discharge in 2001 at the Monywa station (effective rainfall means rainfall minus evapotranspiration)

Table 2: Calibrated parameter values for the Monywa station and the Myitkyina station

n_s ($m^{-1/3}s$)	k_a (m/s)	d_a (m)	d_c (m)	β (-)
The Myitkyina station				
0.703	0.018	2.085	0.455	9.989
The Monywa station				
n_s ($m^{-1/3}s$)	k_a (m/s)	d_a (m)	d_c (m)	β (-)
1.626	0.013	1.627	0.335	7.497

evapotranspiration at 9 stations were used. The location of each data is shown in Fig. 3. Observed river discharge data at Myithyina and Monywa stations were used to calibrate the model parameters. The Nash

Sutcliffe coefficient in Equation (10) was used as an evaluation criterion for calibration:

$$N_s = 1 - \frac{\sum_{i=1}^{n}\left(Q_{obs,i} - Q_{sim,i}\right)^2}{\sum_{i=1}^{n}\left(Q_{obs,i} - \overline{Q_{obs}}\right)^2} \tag{10}$$

$$\overline{Q_{obs}} = \frac{1}{n}\sum_{i=1}^{n} Q_{obs,i} \tag{11}$$

where $\overline{Q_{obs}}$ is a mean of observed discharges. $Q_{sim,i}$ is simulated daily discharge, and $Q_{obs,i}$ is observed daily discharge for the *i*-th day. The SCE-UA algorithm then searches these parameter sets with the highest Nash Sutcliffe coefficient. The calibrated parameters for the two stations are shown in Table 2. Simulation by 1K-DHM-VI then used the same parameters n_s, k_a, d_a, d_c and β as those in Table 2, and the parameter k_0 set to 0.000001.

Results and discussions

The observed discharge and the simulated discharge using 1K-DHM and 1K-DHM-VI at the

Myitkyina station in 2001 are shown in Fig. 4. Fig. 4 indicates that 1K-DHM/1K-DHM-VI reproduced general flow regimes in a rainy season; however, flow regimes in the beginning of a rainy season and quick response of river discharge to rainfall in the dry season (from January through the end of May) are not represented in 1K-DHM. 1K-DHM gives all rain water into subsurface soil layers. Therefore, 1K-DHM is difficult to explain quick response to rainfall in the dry season.

On the other hand, simulated flow regimes by 1K-DHM-VI well reproduced observed river discharge in the beginning of a rainy season. In addition, 1K-DHM-VI also well reproduced sensitive response to rainfall in the dry seasonThis is attributed to the effect of the Hortonian overland flow.

The calibrated result for the Monywa station in 2001 is shown in Fig. 5. It shows that flow regimes simulated by 1K-DHM reproduced general flow regimes in observed in the Monywa station. However, simulated discharge using 1K-DHM does not the responses of observed discharge to high rainfall that occurred in May. 1K-DHM-VI reproduced the quick response of river discharge for the rainfall event. Since calibration of all the six parameters by using 1K-DHM-VI was not conducted, the value of flood discharge is possibly improved by the model calibration.

Figures 4 and 5 also indicated that the model is unable to explain constant base flow in the dry season.

In particular, base flow in the dry season at the Myitkyina station is largely underestimated. Chavoshin *et al*. (2007) showed the same tendency, and MODIS satellite data in Chavoshin *et al*. (2007) shows that base flow in the day season at Myitkyina station is affected by snow melt in an upstream area of the station. Moreover, potential evapotranspiration was directly used as evapotranspiration e, thus e in the dry season is overestimated, which contributed to almost no flow in the dry season in the model.

Conclusions

The DHM-KWSS model structure well represents flow regimes of river basins with temperate climatic conditions; however, it is difficult to explain those in arid basins. To enhance the transferability of DHM-KWSS models for arid basins, this study developed 1K-DHM-VI that added the Hortonian overland flow component on 1K-DHM. Application of 1K-DHM-VI to the Ayeyarwady basin showed the improved reproducibility of sensitive responses of river discharge to rainfall in the beginning of a rainy season and in the dry season. The next step will calibrate all the six parameters for one event and validate the results for other events.

Moreover, the application results also clarified that much more improvement of estimation of actual evapotranspiration and consideration of snow melting are required. Future research must also focus on estimation of meteorological force to capture hydrologic processes in arid areas.

References

Chavoshian, A., Ishidaira, H., Takeuchi, K. and Yoshitani, J., 2007, Hydrological Modeling of Large-scale Ungauged Basin Case Study of Ayeyarwady (Irrawaddy) Basin, Myanmar, HRSD UNESCO-IHP Southeast Asia and the Pacific.

Duan, Q., Sorooshian, S., Gupta, V.K., 1994, Optimal use of the SCE-UA global optimization method for calibrating watershed models, Journal of Hydrology, 158:265-284.

Hunukumbura, P.B., Tachikawa, Y., Shiiba, M., 2012, Distributed hydrological model transferability across basins with different physio-climatic characteristics. Hydrological Processes, 26 (6), 793-808.

Kojiri, T., Hamaguchi, T., Ode, M., 2008. Assessment of global warning impacts on water resources and ecology of a river basin in Japan. Journal of Hydro-environmental Research, 1, 164-175.

Lee, G., Tachikawa, Y., Takara, K., Identification of model structural stability through comparison of hydrologic models. Annual Journal of Hydraulic Engineering, Vol. 51, 49-54, 2007.

Sayama, T., Tachikawa Y., Takara, K., and Ichikawa, Y., 2006. Distributed rainfall-runoff analysis in a flow regulated basin having multiple multi-purpose dams. IAHS Publication, 303, 371-381.

Sayama, T., McDonnell, J.J., 2009. A new time-space accounting scheme to predict stream water residence time and hydrograph source components at the watershed scale, Water Resources Research, 45: W07401, doi:10.1029/2008WR007549.

Sunmin, K., Tachikawa, Y., Nakakita, E., Yorozu, K.., Shiiba, M., 2011. Climate change impact on river flow of the Tone river basin, Japan. Journal of Japan Society of Civil Engineers, Ser. B1 (Hydraulic Engineering), 67 (4), 85-90.

Tachikawa, Y. and Tanaka, T., 2013. http://hywr.kuciv.kyoto-u.ac.jp/products/1K-DHM/1K-DHM.html [Accessed 15 November 2014].

Takasao, T., Shiiba, M., 1988. Incorporation of the effect of concentration of flow into the kinematic wave equations and its applications to runoff system lumping, Journal of Hydrology, 102, 301-322.

Tanaka, T., Tachikawa, Y., 2014, Testing the applicability of a kinematic wave-based distributed hydrologic model in two climatically contrasting catchments, Hydrological Sciences Journal. (this issue)

The Kansai Electric Power Company, 2008, Study Report in Hydrology in Myanmar, 12.