Hydrological forecasting – Prévisions hydrologiques (Proceedings of the Oxford Symposium, April 1980; Actes du Colloque d'Oxford, avril 1980): IAHS–AISH Publ. no. 129.

The Xinanjiang model

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Abstract. This is a conceptual model with distributed parameters, applicable to large basins in the humid regions of China for rainstorm flood forecasting. Its main features are as follows: the concept of runoff formation at natural storage is used for deducting losses, and a high order linear system solution of the Muskingum method is used for flood routing in channels. The model is simple, involving about nine parameters.

Le modèle de Xinanjiang

Résumé. C'est un modèle conceptuel à paramètres distribués, s'appliquant aux grands bassins des régions humides de la Chine pour la prévision de crues. Voilà ses principales caractéristiques: le concept du ruissellement qui est formé au moment où la capacité d'infiltration naturelle du sol est complètement satisfaisante, est utilisé pour déterminer les pertes à déduire, et une solution linéaire d'ordre élevé de la méthode de Muskingum est utilisé pour calculer la propagation des crues dans le lit du canal. Le modèle est simple, il comporte d'environ neuf paramètres.

INTRODUCTION

A stormwater model called Xinanjiang has been established for storm flood forecasting in large basins of the humid regions of China. The concept of runoff formation at natural storage is used for deducting losses, and a high order linear system solution of the Muskingum method is used for flood routing in channels.

THE MODEL OF RUNOFF FORMATION AT NATURAL STORAGE

The control conditions for runoff formation in humid and arid regions are different, and so are the models established for them (East China College of Hydraulic Engineering, 1977). The condition of runoff yield in humid regions is that after the soil moisture content reaches its field capacity, the total rainfall is runoff, the part in excess of infiltration being the surface runoff, and the part infiltrated being the groundwater runoff. In arid regions the natural storage can never be reached, and the condition of runoff yield is that the rainfall intensity exceeds the infiltration rate; all runoff produced is surface runoff. We term the former the runoff formation at natural storage, and the latter the runoff formation in excess of infiltration.

Using the following notation:

P, rainfall in a certain period;
E, evapotranspiration in that period;
D, soil moisture deficit at the beginning of that period;
R, total runoff produced by P, in which surface runoff is RS, interflow is RI and groundwater runoff is RG;
FCA and FCB, final constant infiltration rate of soil zone A and B respectively;
FA, surface infiltration rate.

The runoff formation at natural storage is illustrated by Fig. 1(a). If

$$P-E<D, \text{ then } R=0$$

if $$P-E>D, \text{ then } R=P-E-D$$

Let T be a specified time interval during the runoff yield. If

$$
\begin{aligned}
& (P-E)/T < FCB, \text{ then } RG/T=(P-E)/T, \qquad RI=RS=0 \\
\text{if} \quad & (P-E)/T > FCB, \text{ then } R=P-E, \qquad RG/T=FCB \\
& (RI+RS)/T=(P-E)/T-FCB
\end{aligned}
\tag{1}
$$

That is, when $P-E$ and D are known, it is possible to evaluate R, irrespective of rainfall intensity (or T). By knowing the constant FCB, the groundwater runoff can be separated from the total runoff.

The model of runoff formation in excess of infiltration is shown in Fig. 1(b). When

$$
\begin{aligned}
(P-E)/T < FA, \qquad & RS/T=0 \\
(P-E)/T > FA, \qquad & RS/T=(P-E)/T-FA
\end{aligned}
\tag{2}
$$

RS can be obtained provided that rainfall intensity (or T) and FA are known. FA is not a constant.

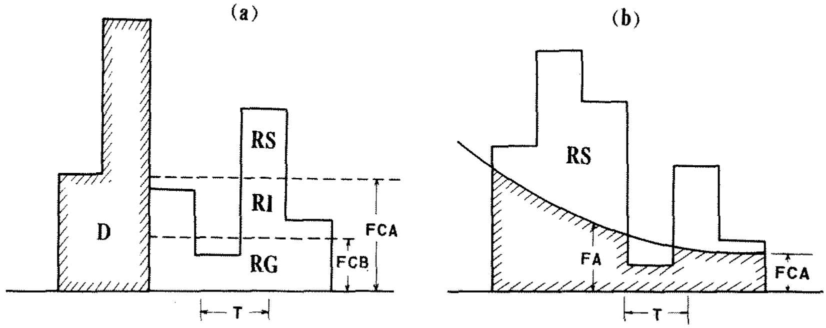

FIGURE 1. Two types of runoff formation.

In the humid regions of China, the model of runoff formation at natural storage has been successfully used (Zhao Ren-jun and Zhuang Yi-ling, 1963) in more than 100 basins. Here water balance computation by equation (5) is applied to evaluate R. The time interval is often taken as a day. In arid regions, the model of runoff formation in excess of infiltration must be used. In using an infiltration curve, the time interval must be small, say 5 min, so as to match the rainfall intensity variations.

The cause of the difference lies in the properties of soil cover and vegetation. In humid regions with rich vegetation and well developed soil zone A, the value of FA is generally far greater than 60 mm/h and the intensity of ordinary rainfall would not exceed it. Besides, the soil moisture deficit in humid regions is not large, generally within a decimetre, and can be recharged easily.

Under dry soil conditions and high intensity rainfall, it may be expected that runoff in excess of infiltration would occur. But in our experience, this is rarely observed in humid regions, probably due to the rapid change of the infiltration curve during the first one or two hours.

While in humid regions surface runoff is often small, the interflow may be very large. Usually the direct runoff consists of 70–80 per cent of the total runoff. The shape of the hydrograph is then flattened. We may see from the hydrograph analysis that the ratio of flood peak to flood amount in arid regions may be 10 times as large as that of humid regions.

In regions of runoff formation at natural storage the problem of the runoff producing area is very important. Because of the non-uniform distribution of soil

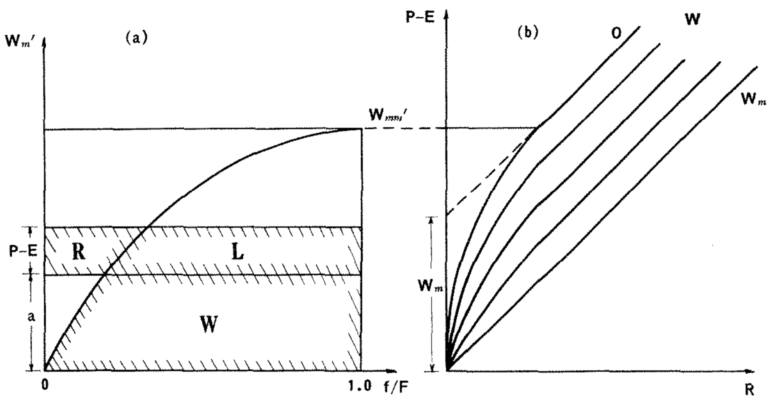

FIGURE 2. Storage capacity curve and rainfall–runoff diagram.

moisture deficit D over the basin, the distribution of runoff yield R is uneven. Because of the non-uniform distribution of FCA, the distribution of RS is uneven.

The storage capacity curve is adopted in the Xinanjiang model to accommodate the non-uniformity of the D distribution. In Fig. 2(a), the ordinate W_m' indicates the storage capacity at a certain point in the basin. Storage capacity means the maximum soil moisture deficit. Abscissa f denotes the partial area of the basin, while F is the total area. The curve represents the area with storage capacity equal to or smaller than a certain value of W_m'.

Making allowance for the horizontal motion of the soil moisture in storage in the basin, we may assume that the distribution of actual storage in the basin W will be characterized by a horizontal line a in Fig. 2(a). Let L be the loss of P, t be the time, then

$$\begin{aligned} L &= P - E - R \\ \mathrm{d}W &= L\,\mathrm{d}t \end{aligned} \tag{3}$$

We fit the storage capacity curve as:

$$f/F = 1 - (1 - W_m'/W_{mm}')^b$$

and then

$$(1 + b)\,W_m = W_{mm}'$$

$$a = W_{mm}'\,[1 - (1 - W/W_m)^{1/(1+b)}]$$

If

$$P - E + a < W_{mm}'$$

then $R = P - E - W_m + W + W_m\,[1 - (P - E + a)/W_{mm}']^{1+b}$

If

$$P - E + a \geqslant W_{mm}'$$

then $R = P - E - (W_m - W)$ (4)

The rainfall–runoff relationship based on equation (4) is shown in Fig. 2(b). If we take the values of P and R of each flood event from the observed data, and make use of the computed values of W, we can calibrate the model.

From equations (3) and (4)

$$P - E - R = W_2 - W_1$$
$$R = f(P - E, W_1) \tag{5}$$

we see, if the values of E are available, the values of R and W could be solved successively by the time interval Δt. Any convenient value may be chosen for Δt. The starting point of the computation may be chosen as $W_1 = W_m$.

In separating groundwater runoff, we use the following criteria:

(1) on the runoff producing area, using equation (1) while $R > 0$,
(2) on the ineffective area, $R = 0$.

From Fig. 2(a)

$$\text{runoff producing area} = R/(P - E)$$
$$\text{ineffective area} = L/(P - E) \tag{6}$$

In China, the regions suitable for modelling at natural storage may be outlined roughly by the following criteria: yearly precipitation > 600 mm and yearly coefficient of runoff > 0.4. The region is extensive.

THE APPLICATION OF THE MUSKINGUM METHOD

The Muskingum method may be considered as a kind of linear system solution. It is a finite difference solution in algebraic form. The physical meanings of its parameters have been clarified as follows (East China College of Hydraulic Engineering, 1977):

$$W = KQ' = K\,[xI + (1 - x)O]$$

$Q' = Q_0$, discharge at steady flow under certain W,

$$x = x_1 - l/2L$$

where l = specific length (Kalinin and Milyukov, 1958), L = length of the reach, and x_1 = coefficient of the wedge shape, being equal to 1/2 when discharge varies linearly along the reach.

$$l = \frac{Q_0}{i_0}\left(\frac{\mathrm{d}H}{\mathrm{d}Q}\right)_0 \doteqdot \frac{\bar{H}}{i_0}$$
$$K = L/\omega_0 \tag{7}$$

ω_0 is wave velocity at steady flow.

These equations serve as an adequate and convenient way to evaluate the parameters.

In solving the linear system problem by the finite difference method, certain linear conditions must be satisfied: (1) discharge varies linearly in time interval Δt at any cross section; (2) discharge varies linearly along the river reach at any time in consideration. So we must take $\Delta t \doteqdot K$ and modify the hydrograph to a broken line. That is, the unit inflow is a triangle with base length $2\Delta t$. The ordinary routing equation including coefficients C_0, C_1 and C_2 conforms to the above conditions.

If Δt is taken far smaller or greater than K, the linear conditions mentioned above would be violated, C_0 or C_2 would take considerable negative values, and a 'negative response' or overestimate would result. In the case of a long reach, in order to satisfy $\Delta t \doteqdot K$, we must use the method of successive routing through sub-reaches, identical to a high order linear system solution. The parameters for each sub-reach can be derived from equation (7), as l and ω_0 will not change when L changes.

This linear system is equivalent to that of Nash (1960), and gives results of the same accuracy. In view of the simplicity of the Muskingum method, we adopt it for the model.

THE XINANJIANG BASIN MODEL

The basin is divided into sub-basins. The outflow from each sub-basin is computed by way of the flow chart shown in Fig. 3. Then they are combined by flood routing in channels to get the basin outlet hydrograph.

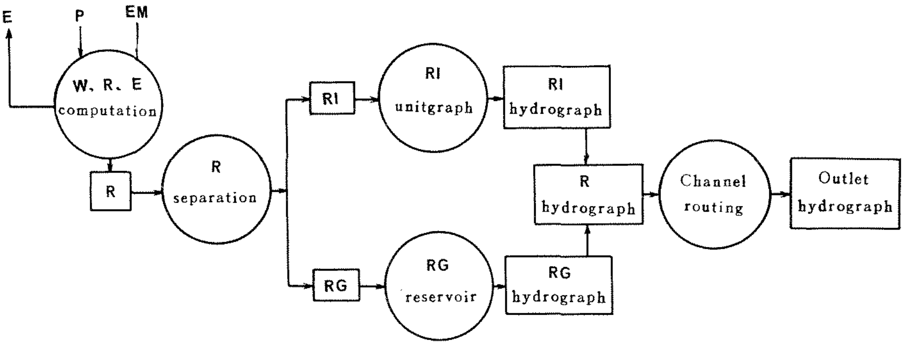

FIGURE 3. Flow chart of Xinanjiang model, for a single sub-basin.

Computation of runoff yield

This is done by equations (5) and (6). Three parameters need to be determined: W_m, b, FCB.

W_m is a climatic factor, denoting the humidity of the basin. In humid regions of south China where annual precipitation amounts to 2000 mm or more, W_m equals 80–90 mm, while in humid regions of north China with annual precipitation 600 mm, W_m equals 140–150 mm. The regional variation is obvious.

Evapotranspiration is often computed by a model comprising three layers of soil moisture capacity. For the upper layer, capacity = 10–20 mm, $E = E_m$; for the middle layer, capacity = 60–100 mm, $E = E_m \cdot W/W_m$; for the lower layer, capacity = 20–70 mm, $E = C \cdot E_m$, $C = 0.10$–0.20.

We usually take the observed values of water evaporation from an E601 pan (area = 3000 cm^2 sunken pan) as E_m; allowing for some corrections $E_m = Z \cdot \text{E601}$. The main factor determining Z is the difference in elevation between the pan station and the basin, as evaporation varies considerably with height.

b is a topographic factor, denoting the non-uniformity of distribution of the soil moisture storage. We take $b = 0.2$–0.4 for mountainous and hilly areas. After the total basin has been charged to its natural storage, b will be ineffective to the function. So b is insensitive for large floods.

Separation of groundwater on the hydrograph by the usual method corresponds to the value of FCB in this model, and is used for evaluating and calibrating the parameter FCB. It is not easy to handle this parameter as the error is comparatively large.

Computation of runoff distribution on sub-basins

For direct runoff the empirical unit graph derived from a representative basin is used. An adequate mathematical expression for the unit graph has not been found. This is a weak point of the model. For groundwater runoff a linear reservoir is used for routing. $W_g = B \cdot Q_g$. Such a simple model can only fit the data at the beginning of the recession.

Flood routing in channels

Parameters K and x are determined by equation (7). The nonlinear effect is often small during high waters.

On the whole, in order to establish the basic structure of the model, it is adequate to use nine parameters: Z, W_m, b, FCB, B, K, x, and two parameters for the unit graph of direct runoff on the sub-basins. Except Z and b, all the parameters can be first estimated directly from observed hydrological data, and then calibrated in the computation routine.

Some parameters are insensitive, such as W_m, b. Some can be easily determined, such as Z, B, x. So there are only three or four parameters which need be estimated carefully. The model can be used without much difficulty.

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