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NUMERICAL SIMULATION OF THE RAINFALL-RUNOFF PROCESS ON A DAILY BASIS

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A digital model has been developed for the simulation of the rainfall - runoff process of rural watersheds. Input data are daily values of precipitation and temperature together with mean monthly potential evapotranspiration. The model produces daily values of streamflow as well as information on the time variation of the soil moisture content. In all, ten model parameters have to be identified, seven of which have a major influence on the performance of the model.
The model operates by accounting continuously for the moisture content in four different and mutually interrelated storages representing physical elements in the watershed.
It has been applied to three different Danish watersheds. Several statistical measures of accuracy have been utilized for a quantitative evaluation of the simulation results. The simulations demonstrate that the main shortcomings of the model are due to the lack of a procedure accounting for frozen ground during extended periods of frost, which could improve some of the simulation results during winter and spring.

The main objective of this study has been to develop a numerical model of the rainfall-runoff process which might, at least for Danish and similar conditions, be considered a fair compromise between input data requirements and the complexity of the model, on the one hand, and acceptable simulation results on the other. As regards the input data requirements, this means that the model operates on the basis of daily values of precipitation and temperature together with mean monthly values of potential evapotranspiration. On this basis the model produces as a main result mean daily values of streamflow as well as information on other elements of the land phase of the hydrologic cycle, such as, for example, the temporal variation of the soil moisture content.

A significant difference between the present model and most other digital models is the way in which the infiltration is determined. In most other models the infiltration is calculated directly by use of a theoretical formula, e.g. Horton's formula. In the present model, however, the amount of infiltrating water is obtained indirectly by a procedure proposed by Marelius (1970). According to this, the infiltration is determined as a residual of the net rainfall (i.e. rainfall diminished by evaporation losses) after subtraction of an empirically obtained amount of direct overland runoff to the stream. This indirect procedure has been chosen partly because theoretical infiltration formulae are not felt to be completely satisfactory for the description of the infiltration process of the watershed as a whole and partly because it has been an objective that the identification of the model parameters should be possible only on the basis of rainfall records and observed streamflow hydrographs.

The model has been applied to three rather different Danish rural watersheds. After presenting the model, some of the simulation results for these watersheds are given and compared with the directly observed streamflow.

THE MODEL

General description

Fig. 1 is a diagram of the structure of the model. As it is seen, it has been attempted to make a simplified imitation of the land phase of the hydrologic cycle. The model operates with four different types of storages interrelated as shown in Fig. 1. The part of the precipitation which has to pass through the snow storage is controlled by the temperature conditions. Moisture intercepted on the vegetation as well as water trapped in depressions and in the uppermost cultivated part of the ground is represented as surface storage. U^* denotes an upper limit of the amount of water in surface storage. The soil moisture in the root zone (i.e. a soil layer below the surface from which losses through evapotranspiration still occur) is represented as lower zone storage. L^* denotes an upper limit of the amount of water in this storage.

Rain and melted snow are first subject to the operation of the surface storage. The amount of water U in surface storage is continuously (i.e. on a daily basis) diminished by evaporative consumption as well as by more or less horizontal leakage (interflow) owing to relatively large horizontal permeabilities in the

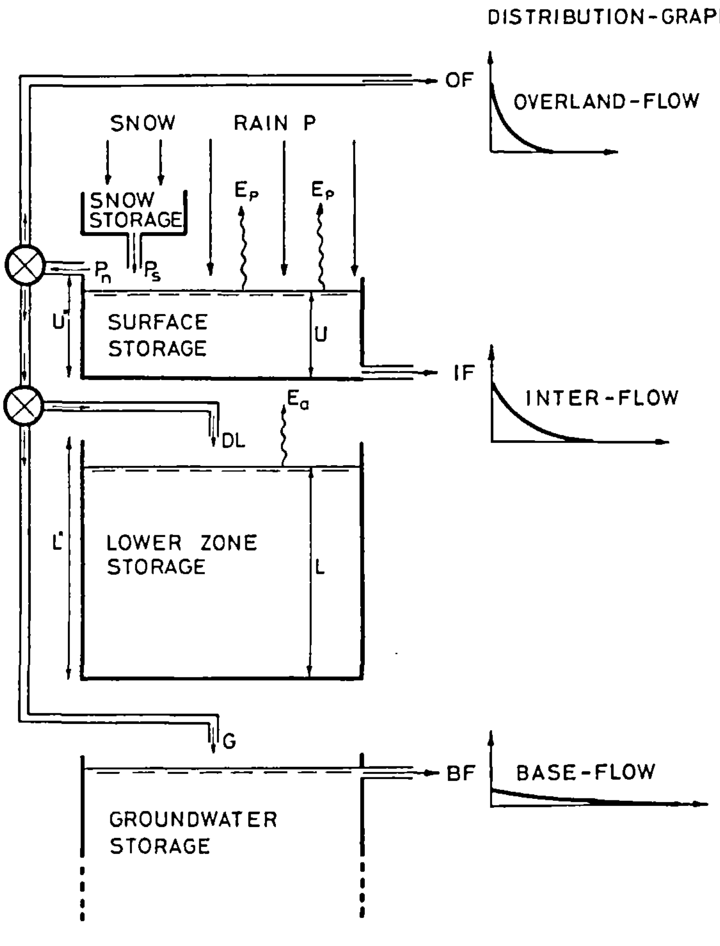

Fig. 1.
Structure of the rainfall-runoff model.

uppermost cultivated part of the soil. When it is full, some of the excess water from the surface storage enters the stream as overland flow, whereas the remainder is diverted as infiltration into lower zone storage and groundwater storage. Moisture in the lower zone storage is subject to a consumptive loss from transpiration, and the moisture content controls the part of infiltrating water which enters the groundwater storage.

Groundwater storage is continuously drained to the stream as base flow. Corresponding to the action of simple linear reservoirs, the outflows from the various types of storages in the form of overland flow, interflow and base flow are routed to the stream according to different exponential lag or distribution functions for the particular type of flow considered. By adding up the different kinds of contributing flows we obtain as a result a continuous streamflow hydrograph.

The quantitative relations governing the operation of and the interrelationship between the various storages are described in the following sections.

Snow storage

When the mean daily temperature T is below the freezing point, precipitation is assumed to fall as snow. It is accumulated in the snow storage until melting conditions occur. The mean daily air temperature is obtained as $T = (T_{min} + T_{max})/2$. When T is above the freezing point, the snow remaining in storage is assumed to release a daily amount of melting water, $P_s = C_s T$, proportional to the temperature T in Centigrade. The parameter C_s for a specific watershed is estimated on the basis of the observed streamflow hydrographs corresponding to snow-melting situations.

Surface storage

In short, the action of this storage consists in receiving rain and melted snow and in regulating overland flow as well as evaporative losses and interflow.

As long as any water is present in surface storage the moisture content U is continuously reduced by potential evapotranspiration and interflow. The daily amount of potential evapotranspiration E_p is obtained directly on the basis of the monthly values observed. The daily amount of moisture, IF, contributing to interflow is assumed proportional to U and to vary linearly with the relative moisture content L/L^* of the lower zone storage

$$IF = \begin{cases} C_{IF} \dfrac{L/L^* - C_{L1}}{1 - C_{L1}} U & \text{for } L/L^* > C_{L1} \\ 0 & \text{for } L/L^* \leqq C_{L1} \end{cases} \qquad (1)$$

L denotes the moisture content of the lower zone storage while the parameters C_{IF} and C_{L1} are both positive and dimensionless constants smaller than unity. IF, as obtained from eq. (1) for a particular day, is routed to the stream as interflow during the subsequent days according to an exponential weighting function $\frac{1}{K_I} e^{-t/K_I}$ $(t = 0,1,2,\ldots)$. This corresponds to the action of a simple linear reservoir having a time constant K_I. For the watersheds considered in this study K_I is approximately 3 days. C_{IF}, C_{L1} and K_I have to be estimated from available streamflow hydrographs more or less by trial and error.

The surface storage has to be filled, i.e. $U \geqq U^*$, before any excess water P_n occurs. Thus the maximum capacity U^* of the surface storage can be estimated after a long and dry period as the amount of net rainfall that has to occur before any overland flow takes place.

When the surface storage spills, i.e. when $U \geqq U^*$, the excess water P_n gives rise to overland flow as well as to infiltration. OF denotes that part of P_n which contributes to overland flow. It is assumed to be proportional to P_n and to vary linearly with the relative moisture content L/L^* of the lower zone storage

$$OF = \begin{cases} C_{OF} \dfrac{L/L^* - C_{L2}}{1 - C_{L2}} P_n & \text{for } L/L^* > C_{L2} \\ 0 & \text{for } L/L^* \leqq C_{L2} \end{cases} \tag{2}$$

The parameters C_{OF} and C_{L2} are both positive and dimensionless constants smaller than unity. C_{OF} is estimated as the ratio between the accumulated amount of overland flow and the excess rainfall in connection with such runoff events during the winter ($L \simeq L^*$) for which the rainfall gives rise to rapid and large increases in the streamflow. C_{L2}, on the other hand, is estimated on the basis of such situations where even very heavy rainfall does not give rise to an increase in the streamflow. OF, as obtained from eq. (2) for a particular spilling situation, is routed to the stream as overland flow during the subsequent days according to an exponential weighting function $\frac{1}{K_O} e^{-t/K_O}$ $(t = 0,1,2,\ldots)$. This corresponds to the action of a simple linear reservoir having a time constant K_O. K_O is estimated from streamflow situations identical to those giving rise to the estimate of C_{OF}. For the watersheds considered in this study K_O is approximately 2 days.

Lower zone storage

That part of the net rainfall excess P_n that does not run off as overland flow infiltrates into the lower zone storage representing the root zone.

One part DL of the amount of infiltration (P_n–OF) is assumed to increase the moisture content L in lower zone storage. DL is set proportional to the deficit of moisture ($1-L/L^*$) in lower zone storage

$$DL = (P_n - OF)\,(1 - L/L^*) \tag{3}$$

The rest of the amount of infiltrating moisture, G, is assumed to percolate deeper into groundwater storage

$$G = (P_n - OF)\, L/L^* \tag{4}$$

As previously mentioned, moisture in the lower zone storage is subject to a consumptive transpiration loss. Evapotranspiration demands are at first attempted to be met at the potential rate from the surface storage. If the moisture content U in the surface storage is too small to fulfil completely these requirements, a fraction of the rest is assumed to be withdrawn at an actual rate by root activity from the lower zone storage. The actual evapotranspiration E_a is put equal to the potential evapotranspiration E_p multiplied by the relative water content L/L^* in lower zone storage for that particular day

$$E_a = E_p \cdot L/L^* \tag{5}$$

The maximum moisture content L^* in the lower zone storage may be estimated from a knowledge of the field capacity within the watershed.

Groundwater storage

The build-up of moisture in the groundwater storage is determined by eq. (4). The groundwater storage is assumed to act as a linear reservoir, i.e. the outflow is proportional to the water content in the storage. This is equivalent to an assumption that the outflow follows an exponential recession curve and thus the daily amount of water routed to the stream as base flow BF may be obtained by the following expression

$$BF = BF_O \cdot e^{-1/K_B} + G(1-e^{-1/K_B}) \tag{6}$$

where BF_O is the amount of base flow the day before, K_B is the time constant of the groundwater reservoir and G is the amount of water percolating into groundwater storage on the actual day. The time constant K_B for a given watershed may be estimated on the basis of the recession of the streamflow hydrograph for a long period with negligible amounts of rain.

SIMULATION RESULTS

General on accuracy

Obviously a streamflow model should be able to produce simulation results of an acceptable accuracy, i.e. to respond to different meteorological events as a natural watershed would respond. A procedure for evaluating whether this objective is fulfilled by the model might be based on a comparison between simulated and recorded streamflow records. However, even if such a comparison shows good agreement, it does not ensure that the model simulates correctly the various physical processes occurring in the watershed. For a more complete evaluation of the physical validity of the simulation model it would be desirable to consider measurements of the time variation of other variables in the model, such as, for instance, soil moisture, overland flow, etc. Generally, however, only measurements of streamflow are available to test the qualifications of the model. On the other hand, the larger the number of different watersheds which can be simulated with acceptable accuracy by the model, the greater the confidence that the parameters and model structure utilized have a realistic physical meaning. In this section we will test the ability of the present model to simulate the runoff characteristics of three rather different Danish watersheds.

It is important to point out that besides the adequacy of the model in simulating the physical processes involved, the comparison between simulated and recorded streamflow is influenced by several other factors to be considered more or less as noise in this connection. In short, these are

a) the inaccuracy in the measurement of streamflow as well as in the point measurements of precipitation, potential evapotranspiration and temperature.

b) the adequacy of point measurements of the meteorological data in representing the true spatial mean values over the watershed.

These factors set an upper limit to the attainable degree of correspondence between the simulated and observed stream flow. This limit has nothing to do with the validity of the model in describing the physical processes of the hydrologic cycle within the watershed.

Statistical measures of accuracy

For the purpose of evaluating quantitatively the accuracy of the model as well as for the comparison between different simulations corresponding to different values of the model parameters, we make use of the following statistical measures:

1. For each year i of the considered N years of simulation and observation of streamflow, the correlation coefficients r_i^Y obtained by matching daily simulated streamflow Q'_{ij} $(j = 1, \ldots, 365)$ and daily recorded streamflow Q_{ij} are obtained by

$$r_i^Y = \frac{\sum\limits_j (Q_{ij}-YQ_i)\ (Q'_{ij}-YQ'_i)}{\sqrt{\sum\limits_j (Q_{ij}-YQ_i)^2 \sum\limits_j (Q'_{ij}-YQ'_i)^2}} \ ; \ (i = 1, \ldots, N) \qquad (7)$$

where

$$YQ_i = \frac{1}{365}\sum_{j=1}^{365} Q_{ij} \text{ and } YQ'_i = \frac{1}{365}\sum_{j=1}^{365} Q'_{ij} \tag{8}$$

2. For each year i is determined the part of the total variance of the daily recorded streamflow Q_{ij} which is described by the model

$$R^2_i = \frac{\sum_j (Q_{ij}-YQ_i)^2 - \sum_j (Q_{ij}-Q'_{ij})^2}{\sum_j (Q_{ij}-YQ_i)^2}, \ (i = 1, \ldots, N) \tag{9}$$

3. The two measures mentioned above are often used in connection with the analysis of streamflow models (Crawford & Linsley 1966, Porter & McMahon 1971, Nash & Sutcliffe 1970). However, in judging the feasibility of the model in simulating the behaviour of different watersheds, both of them have the following undesirable property: The greater the mean seasonal variation in streamflow, the less these measures tell about the ability of the model to describe the more stochastic variation in the streamflow. To obtain a measure taking account of this, we may separate the mean seasonal variation and consider only the irregular variation about this more or less deterministic element in the streamflow record. By matching for each day j in the year these deviations of the simulated and recorded daily streamflow, the following correlation coefficient r_j^D is obtained

$$r_j^D = \frac{\sum_i (Q_{ij}-DQ_j)\ (Q'_{ij}-DQ'_j)}{\sqrt{\sum_i (Q_{ij}-DQ_j)^2 \sum_i (Q'_{ij}-DQ'_j)^2}}, \ (j = 1, \ldots, 365) \tag{10}$$

where

$$DQ_j = \frac{1}{N}\sum_{i=1}^{N} Q_{ij} \text{ and } DQ'_j = \frac{1}{N}\sum_{i=1}^{N} Q'_{ij} \tag{11}$$

4. The daily values of recorded and simulated streamflow are added to monthly values q_{ij} and q'_{ij}, respectively. Correlation coefficients r_j^M for monthly streamflow analogous to those described above for daily values are calculated by

$$r_j^M = \frac{\sum_i (q_{ij}-\bar{q}_j)\ (q'_{ij}-\bar{q}'_j)}{\sqrt{\sum_i (q_{ij}-\bar{q}_j)^2 \cdot \sum_i (q'_{ij}-\bar{q}'_j)^2}}, \ (j = 1, \ldots, 12) \tag{12}$$

where

$$\bar{q}_j = \frac{1}{N}\sum_{i=1}^{N} q_{ij} \text{ and } \bar{q}'_j = \frac{1}{N}\sum_{N}^{i=1} q'_{ij} \tag{13}$$

Examples of simulations

The simulation model has been applied to three Danish watersheds drained by the following watercourses: Skjern å in western Jutland, Grønå in southern Jutland and Tryggevælde å in eastern Zealand. A summary of the main characteristics of the three watersheds as well as information on the simulation periods are given in Table 1. Maps of the different watersheds as well as the locations

Table 1.

	Skjern Å	Grønå	Tryggevælde Å
Streamflow gauging station	Gl. Alergård	Vintved kanal	Ll. Linde
Watershed area	1056 km^2	579 km^2	131 km^2
Average annual precipitation (appr.)	750 mm	850 mm	610 mm
Average annual runoff (appr.)	415 mm	490 mm	200 mm
Average annual pot. evap. (appr.)	460 mm	400 mm	500 mm
Precipitation stations used	Høgildgård + Give	St. Jyndevad + Gråsten	Tureby + Karise
Evap. stations used	Borris + Studsgård	St. Jyndevad	Højbakkegård (1958–67) Endeslev (1968–71)
Temp. station used	Studsgård	St. Jyndevad	Sorø
Simulation period	1956–68	1960–68	1958–71
Years used for parameter identification	1958 + 1962	1961 + 1964	1961 + 1962
			(In all years except 1961–62 only monthly values of streamflow have been considered)

of the observational stations are shown in Fig. 9. Potential evapotranspiration data have been obtained from pan evaporation observations after correction by suitable seasonal-dependent factors as described by Knudsen (1969).

For each of the three watersheds Table 2 gives the model parameters utilized by the simulations.

In all, ten model parameters have to be identified. Of these, seven have a major influence on the performance of the model, namely the melting constant C_s, the maximum moisture capacities U^* and L^* of the surface and the lower zone storages, respectively, the constant C_{OF} controlling to a large extent the fraction of excess rainfall contributing to overland flow, and finally the time constants K_0, K_I and K_B of the distribution graphs for overland flow, interflow and base flow, respectively. Concerning the remaining three parameters C_{IF}, C_{L1} and C_{L2}, the overall behaviour of the model is less sensitive to changes in their numerical values.

The only significant differences between the Skjern å and the Grønå watersheds are the maximum capacity of the surface storage and the behaviour of the base flow. As indicated by the magnitude of K_B, the base flow recession is much faster in the Grønå watershed.

From Tables 1 and 2 it is seen that the behaviour of the Tryggevælde å watershed differs considerably from the other two watersheds considered. Thus, the large values of C_{OF} and L^* indicate that the groundwater storage in this watershed is rather unimportant as compared with the action of the upper

Table 2.

Parameter		Skjern Å 1056 km²	Grønå 579 km²	Tryggevælde Å 131 km²
Capacities of storages (mm water)	U^*	10 mm	15 mm	15 mm
	L^*	100 mm	100 mm	150 mm
Melting constant C_s		2 mm/°C/day	2 mm/°C/day	2 mm/°C/day
Overland flow	C_{OF}	0.15	0.20	0.60
	C_{L2}	0.7	0.7	0.7
	K_0	2.5 days	2.5 days	1.9 days
Interflow	C_{IF}	0.06	0.04	0.05
	C_{L1}	0	0	0.7
	K_I	3.3 days	3.3 days	2.8 days
Base flow K_B		333 days	83 days	67 days

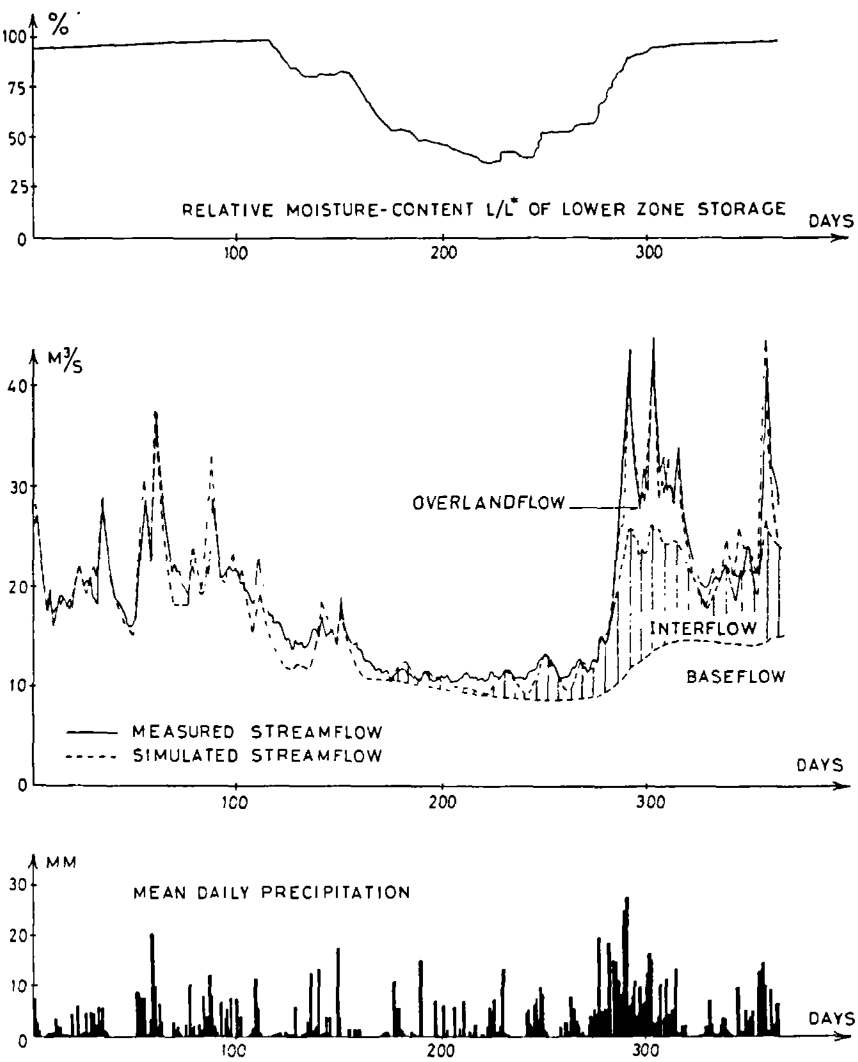

Fig. 2.
Simulation results - Skjern å 1967.

storages. Furthermore, larger actual evapotranspiration is to be expected for the Tryggevælde å watershed because of larger values of the potential evapotranspiration E_p and the maximum capacity L^* of the lower zone storage. The smaller time constants of the distribution graphs are mainly to be attributed to the smaller size of the Tryggevælde å watershed.

For each of the watersheds considered, examples are given for one year in Figs. 2–4 of the observed and simulated streamflow and of the relative moisture content L/L^* in lower zone storage as obtained by the simulation. The various components of the total simulated streamflow, i.e. base flow, interflow and overland flow, are indicated for Skjern å in Fig. 2. This gives a qualitative impression of their relative magnitudes to judge from the model.

Some streamflow situations in winter and in early spring are rather poorly simulated by the model. As shown in Fig. 4, this is the case with, for instance, Tryggevælde å during the period February–March, 1962. This may be attributed to the fact that the model does not account for the influence of frozen

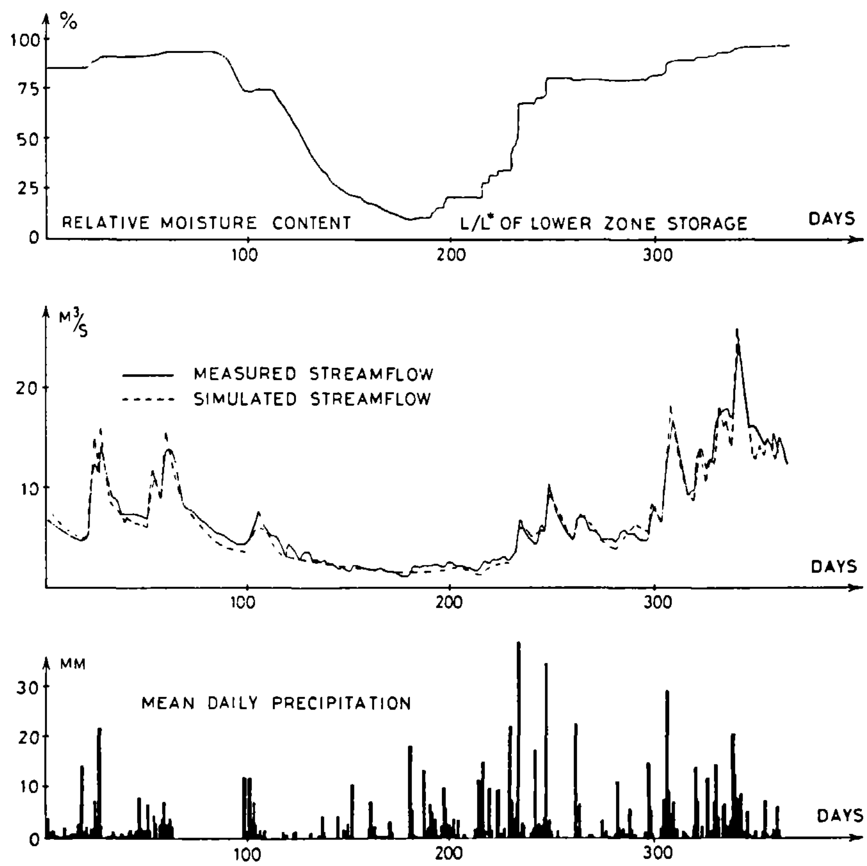

Fig. 3.
Simulation results – Grønå 1960.

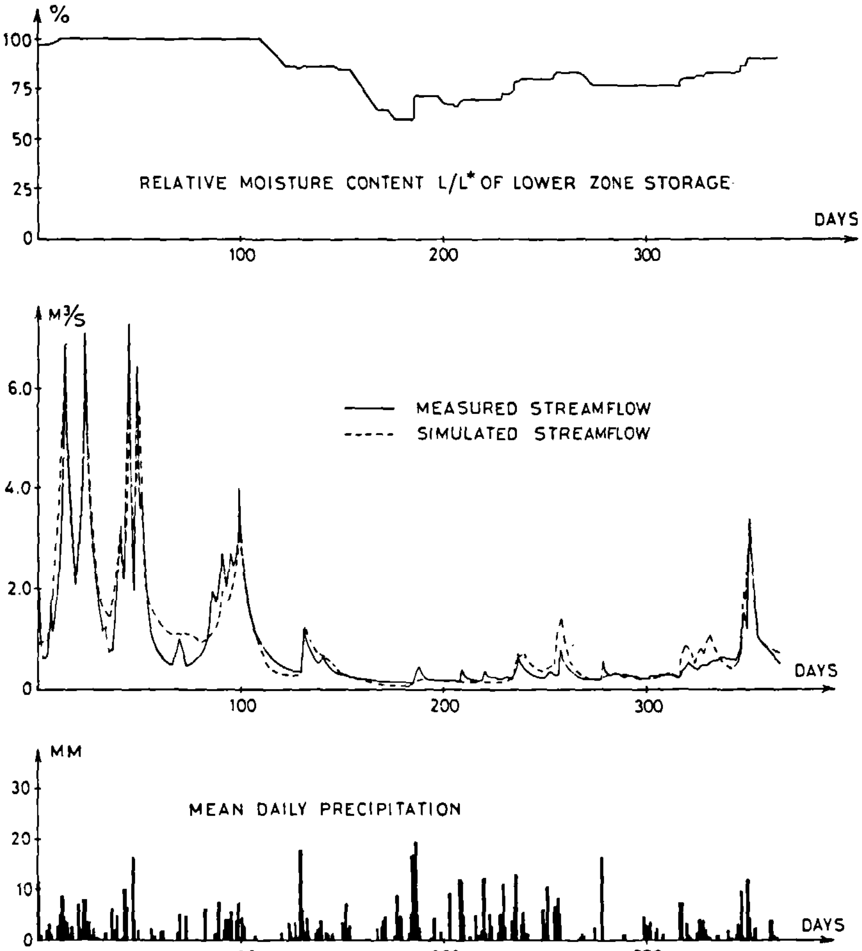

Fig. 4.
Simulation results – Tryggevælde å 1962.

ground. In nature, however, extended periods of frost give a rather complicated runoff pattern mainly because much of the water in the upper storages gets frozen and is thus prevented from moving until released during a subsequent thawing period.

The relative moisture content L/L^* shown in Figs. 2–4 is, of course, to be considered as mean values all over the watershed in question. Qualitatively the time variations are found to correspond very well with published Danish data for the soil moisture variation in the root zone during the growing season

(Aslyng & Kristensen 1958). This gives some confidence that the model is yielding a physically realistic description of the soil moisture conditions.

The statistical measures r_i^Y and R_i^2, as defined for daily values by eqs. (7) and (9), respectively, are given in Table 3 for Skjern å and Grønå for each year in the simulation periods considered. For the years 1963 and 1966, observed streamflow data are lacking for Grønå.

Table 3.

Year i	Skjern Å		Grønå	
	r_i^Y	R_i^2	r_i^Y	R_i^2
1956	.886	.77		
57	.950	.83		
58	.914	.83		
59	.935	.81		
60	.935	.77	.985	.97
61	.944	.84	.970	.92
62	.883	.77	.952	.88
63	.928	.81	–	–
64	.943	.64	.960	.91
65	.932	.81	.946	.89
66	.928	.82	–	–
67	.974	.94	.928	.86
68	.889	.75	.960	.91
Mean	.926	.79	.978	.96

According to Table 3 the model seems to give a better simulation of the Grønå watershed than of the Skjern å watershed.

The mean flows DQ_j and DQ'_j ($j = 1, \ldots, 365$), as obtained from eq. (11) for each day in the year, are shown in the lower part of Figs. 5 and 6 for Skjern å and Grønå, respectively. The graphs of the recorded mean flow DQ_j show the more or less deterministic seasonal variations in the runoff from the two watersheds. The graphs of $e_j = DQ'_j - DQ_j$ may be used to evaluate the ability of the model to simulate these seasonal variations. Besides statistical sampling errors as well as errors in the observation of the streamflow, three major

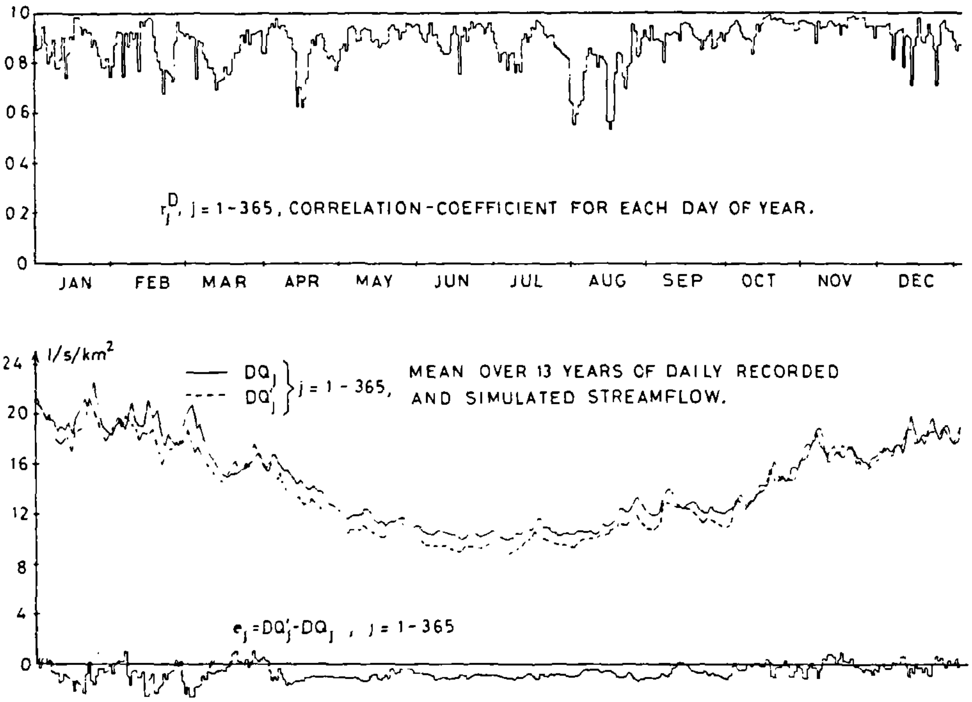

Fig. 5.
13-year mean daily simulated flow – Skjern å 1956–68.

reasons may be given for e_j being different from zero: First and secondly, more or less systematic registration errors in the input data of precipitation and potential evapotranspiration, and thirdly, of course, errors in the model estimates of the actual evapotransporation. Naturally, all of these reasons may come into play. In the case of the Skjern å, it is obvious from Fig. 5 that the simulated flows are systematically less than the recorded flow most of the year. According to Madsen (1972), the winter precipitation at many Danish stations and especially in western Jutland is seriously underestimated due to difficulties in the registration of snowfall amounts. This may explain why e_j for Skjern å is significantly less than zero during the winter. The too low streamflow simulated in Skjern å in the summer period must, however, be attributed to either excessive registrations of potential evapotranspiration or to incorrect model estimates of the actual evapotranspiration, or both.

In the case of Gronå, Fig. 6 indicates that the water balance on a yearly basis is well satisfied. Also here, however, some discrepancies show up in the winter period, whereas the mean runoff pattern in the summer months is simulated quite accurately.

In the upper part of Figs. 5 and 6 are plotted for Skjern å and Grønå, respectively, the calculated correlation coefficients r_j^D, as obtained from eq. (10) for each day of the year. It is seen that according to this measure, the ability of the model to simulate the stochastic variation of the daily runoff pattern is best during the late fall and the winter period. This may be due to the small influence of the rather uncertain evapotranspiration during these periods.

In Fig. 7 are plotted all the series of monthly simulated and recorded streamflows for each of the three watersheds considered. From this we realize the rather different regimes of the three watercourses, Skjern å being to a large extent fed by ground water and Tryggevælde å nearly exclusively by surface water and, finally, Grønå being between the two. The great difference between

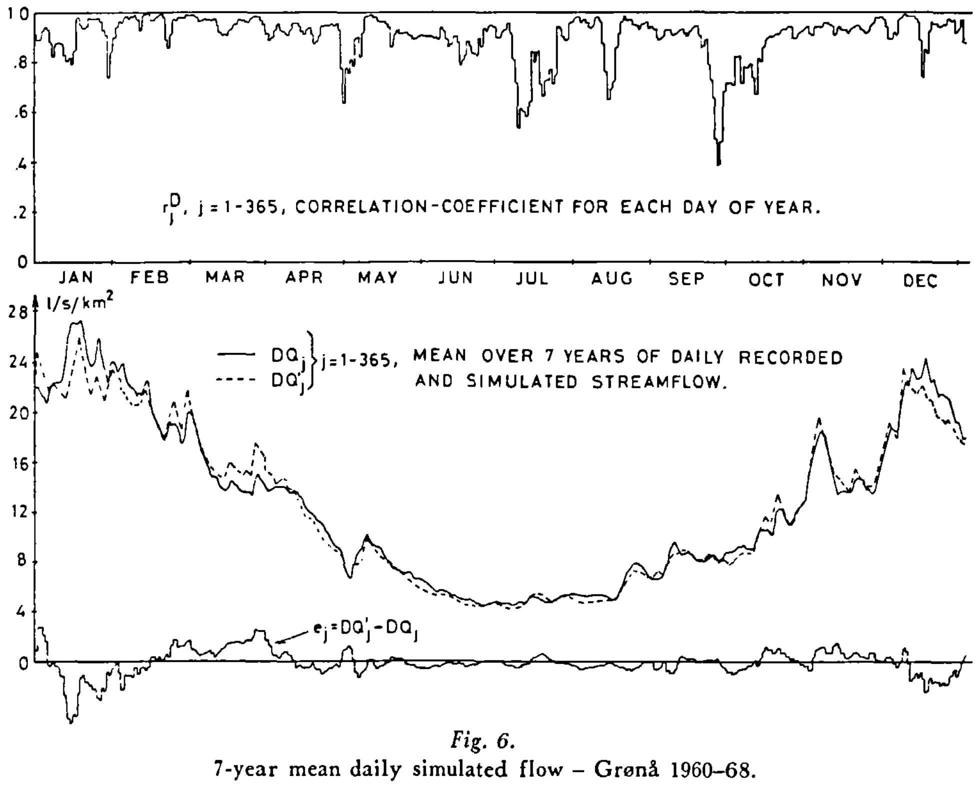

Fig. 6.
7-year mean daily simulated flow – Grønå 1960–68.

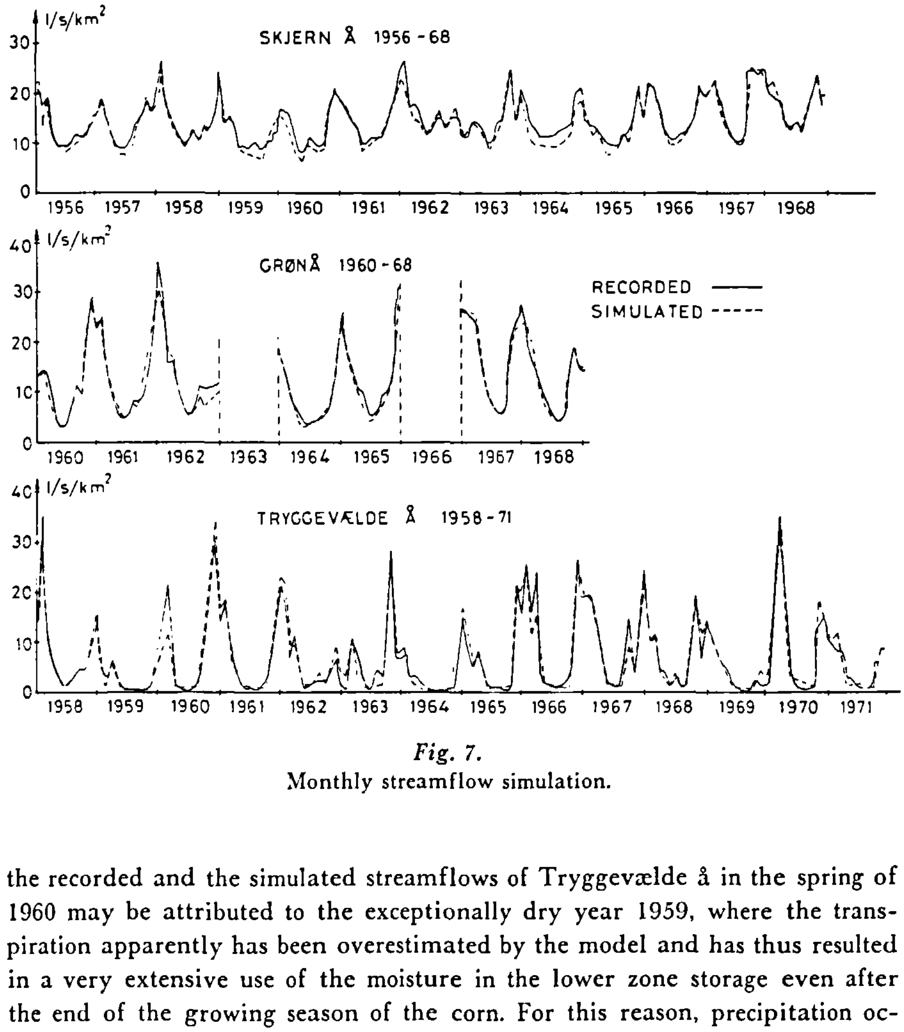

Fig. 7.
Monthly streamflow simulation.

the recorded and the simulated streamflows of Tryggevælde å in the spring of 1960 may be attributed to the exceptionally dry year 1959, where the transpiration apparently has been overestimated by the model and has thus resulted in a very extensive use of the moisture in the lower zone storage even after the end of the growing season of the corn. For this reason, precipitation occurring in the early spring of 1960 in the model has been mainly diverted as infiltration to the lower zone storage and not in sufficient amounts to the stream as overland flow and interflow.

Fig. 8 is similar to Figs. 5 and 6 with the exception that in Fig. 8 monthly values are considered. q_i, q'_j and r_j^M, as calculated from eqs. (13) and (12), respectively, are plotted for all of the three watersheds. The plots in Fig. 8 for Skjern å and Grønå naturally show a great deal of similarity to the corresponding plots based on daily values and commented on previously. Further, it is observed that the simulations for Tryggevælde å are not as satisfactory as for Skjern å and Grønå, especially not as regards the simulation of the

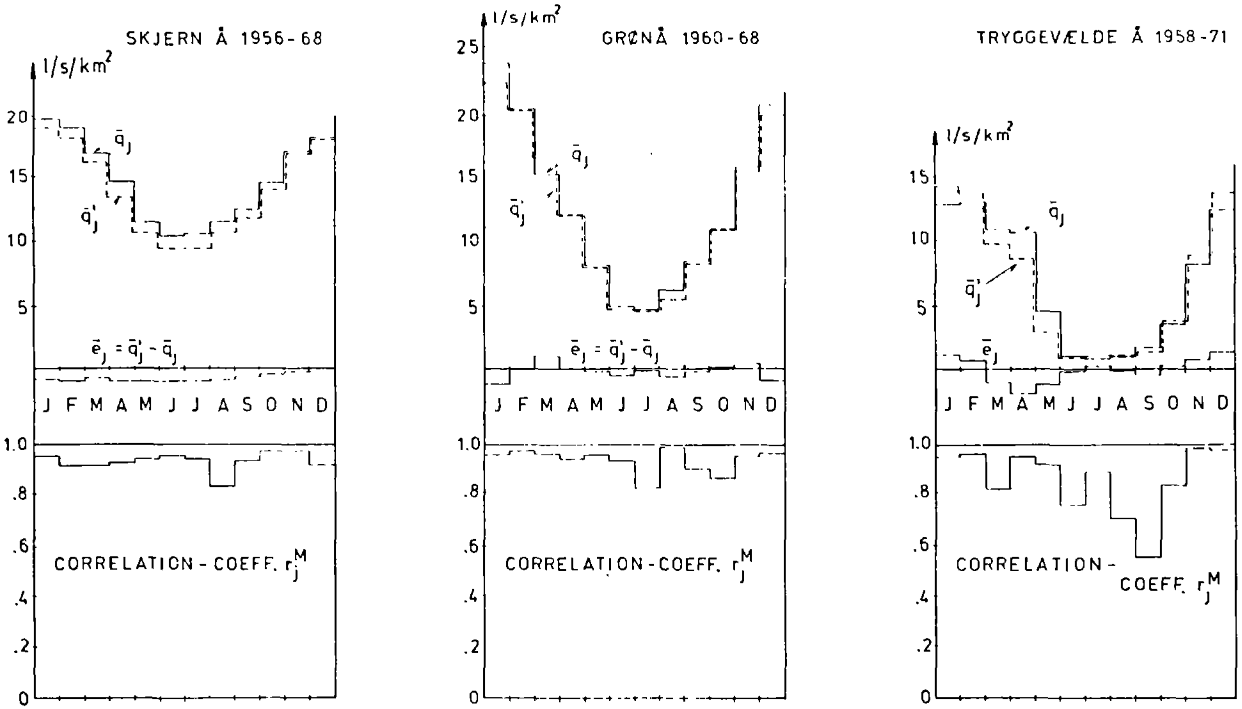

Fig. 8.
Mean over N years of monthly recorded and simulated flow.

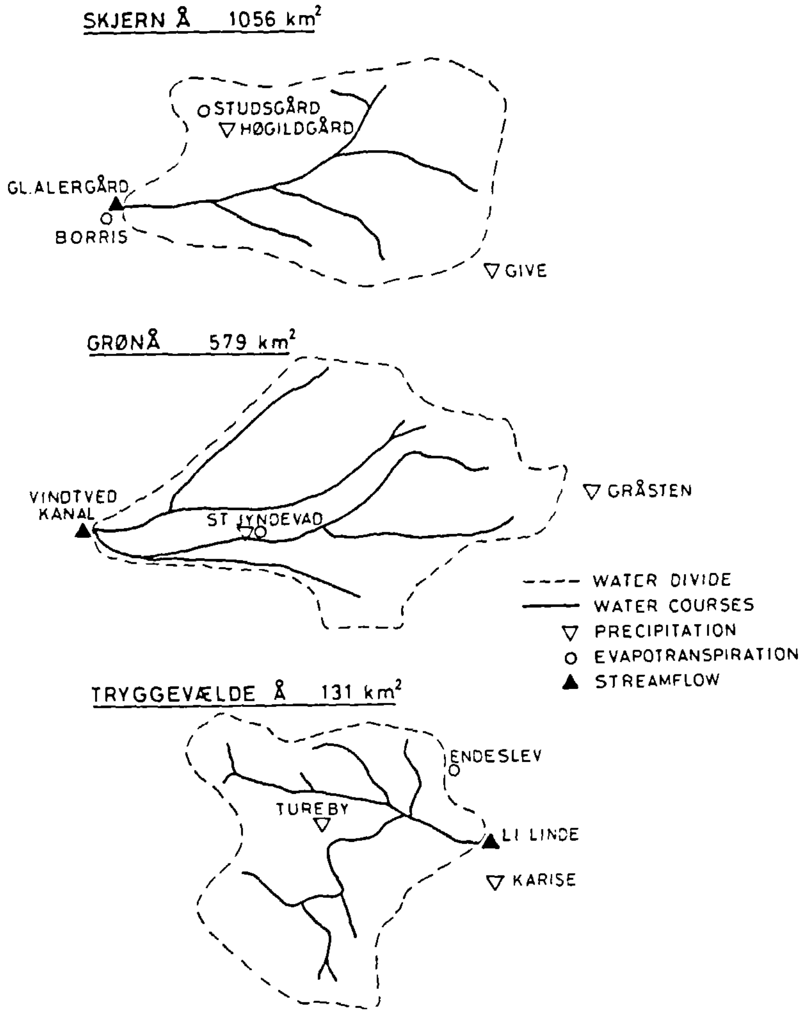

Fig. 9.
Maps of the watersheds and locations of precipitation, evapotranspiration and streamflow stations.

stochastic variations in the monthly streamflow. However, the relatively small values of the correlation coefficients r_j^M in the summer months for Tryggevælde å may probably be attributed to the complete lack of potential evapotranspiration data within the watershed for the period before 1968. Evapotranspiration data relatively far from the watershed (see Table 1) have been utilized for the period 1958–68.

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