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Modelling Water Exchange and Transit Times in Till Basins Using Oxygen-18

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The application of hydrological models to data on conservative tracers can yield information about transit times and storage volumes and may provide an independent test of the model structure.

In this study, the PULSE-model has been modified to simulate conservative tracers. Attempts have been made to describe both short-term and long-term variations in oxygen-18 concentration in three small forested basins.

The performance of the model was considerably increased, when additional storage was introduced in the model. The turnover times were estimated to approximately 7 and 12 months for two of the basins.

Introduction

A conceptual runoff model may reproduce runoff records fairly well, even if the model structure is a poor description of the physical reality. However, if a conceptual model is to be the hydrological basis for hydrochemical simulations, the model has to describe the runoff process, storage volumes and transit times in a reasonable way. This is especially true, if predictions are to be made, which incorporate time-dependent chemical reactions.

Environmental isotopes have been used in a large number of studies concerning processes in the hydrological cycle. These studies yield information about the process of streamflow generation (e.g. Sklash and Farvolden 1979, Rodhe 1981 and 1984) and about transit times (e.g. Eriksson 1963, Dinçer *et al.* 1970, Maloszewski *et al.* 1983). A summary of the application of environmental isotopes to modelling in hydrology was given by Dinçer and Davis (1982).

The application of conceptual hydrochemical models to data on conservative tracers may provide an independent test of the model structure and be of great value when developing the model towards a more physically sound description of runoff. Such an application was performed by Christophersen *et al.* (1985), who attempted to simulate the flow of ^{18}O through the Birkenes basin using the Birkenes model. They concluded that the original reservoir volumes of the model had to be increased in order to obtain a sufficient damping of the input fluctuations in ^{18}O-concentration.

In this paper, the PULSE-model (Bergström *et al.* 1985) is modified to simulate the ^{18}O-concentration of streamwater from the ^{18}O-concentration of precipitation, which can vary considerably between precipitation events.

Model Structure

The basic structure of the modified PULSE-model used in this work is shown in Fig. 1. The flowpaths of this model structure have certain similarities with the "model 3", proposed by Maloszewski *et al.* (1983), but the two models differ in the water balance calculations. Details about the water balance calculations, which are identical to those of the original PULSE-model, were given by Bergström (1976) and Bergström and Sandberg (1983).

Input requirements are daily mean values of temperature, daily amounts and $\delta^{18}O$ of precipitation*), and monthly standard values of potential evapotranspiration. Calculations are performed with a timestep of one day.

Except for the fractionation taking place during snow storage and snowmelt, ^{18}O is regarded in the model as a conservative tracer during water flow in the basin. Fractionation during interception, as reported by Saxena (1986), is thus disregarded.

Snow

The snowpack is treated as homogeneous with respect to ^{18}O. Isotopic stratification due to variation in ^{18}O content between precipitation events as described by Stichler *et al.* (1981) is consequently neglected.

Due to fractionation by melting/freezing, meltwater $\delta^{18}O$ (δ_m) often is depleted as compared with the snowpack (δ_{sp}) (cf. Herrmann *et al.* 1981, and Stichler *et al.* 1981). This is parameterized according to

$$\delta_{sp} - \delta_m = 2e^{-\gamma m/sp} \qquad (\delta \text{ in } {}^0/_{00}, \quad 0<m<sp) \tag{1}$$

* The ^{18}O-concentration of a water sample is expressed as $\delta^{18}O$, i.e. as the relative deviation of the isotopic ratio $^{18}O/^{16}O$ of the sample to that of a standard water (SMOW).

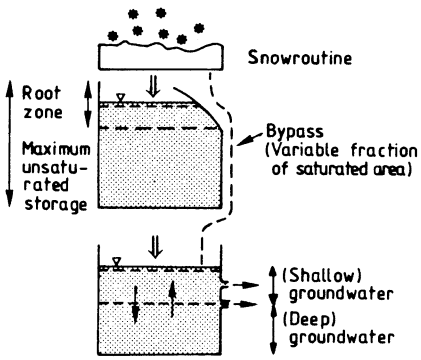


Fig. 1. Schematic flowpaths when simulating ^{18}O by the modified PULSE-model.

where *m* and *sp* are the meltwater volume (in one day) and the water equivalent of the snowpack respectively, and γ is a model parameter.

Molecular exchange with the atmosphere generally enriches the snowpack in ^{18}O, described in the model as

$$\delta_{sp}(t+1) \equiv \delta_{sp}(t) + \frac{\varepsilon}{sp} \tag{2}$$

with ε being approximately 1-2 [‰ mm/day], according to field measurements during meltfree periods (Rodhe unpublished).

Soil Moisture

The soil moisture storage as calculated by the original model corresponds to the amount of water between field capacity and the wilting point. In order to describe the total volumes of water active in the hydrological cycle, an additional volume mainly representing water in the unsaturated zone below the root zone and water held below the wilting point, was introduced. This volume does not affect the water balance computations.

Saxena (1984) was able to trace soil moisture layers, depleted in ^{18}O, for two consecutive melt periods in a till soil in the Kloten area, close to the Buskbäcken basin, studied in this paper. This indicates that water in the unsaturated zone percolates in a manner close to piston-flow, pushing down older water towards the groundwater zone. To describe the flow in the unsaturated zone, this concept was initially used, but it failed, since it results in a mere delay in time without any damping of computed $\delta^{18}O$. Even if piston-flow prevails locally in the unsaturated zone, the groundwater recharge in a basin is a mixture of water with different transit times in the unsaturated zone.

In this application of the modified PULSE-model, each pulse of snowmelt or rain is therefore individually traced through the unsaturated zone, but the outflow is composed of an equal portion of each pulse in the zone. This is equivalent to what is referred to as ideal mixing (as far as transit times are concerned). Evapo-

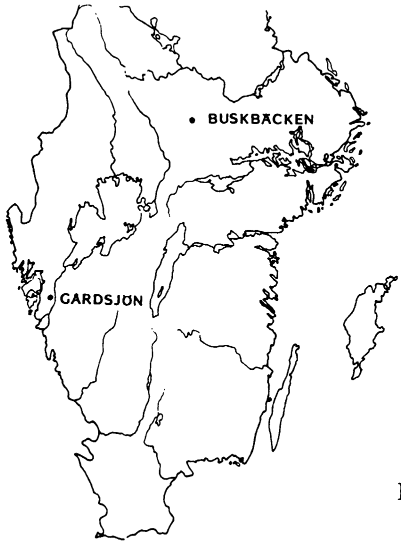


Fig. 2. Location of the sites in the south of Sweden.

transpiration is, however, assumed to take part exclusively in the root zone, thus somewhat distorting the ideal mixing.

Groundwater

Precipitation falling on saturated areas, close to the stream, cannot infiltrate and forms saturation overland flow. This is accounted for by the introduction of a bypass (Fig. 1). A certain fraction of the generated runoff is thus assumed to consist of overland flow on saturated areas. The extent of saturated areas is related to the soil moisture storage in the model and increases with increasing wetness in the area.

A reservoir, representing deep groundwater and other additional water, is incorporated in the model (Fig. 1). The volume of this reservoir is assumed to be constant and does not affect runoff calculations. Each day a portion of the water in the groundwater tank is exchanged with an equal volume from this additional groundwater storage. The fraction of deep groundwater, contributing to streamwater, will consequently rise during the recession after a flow event.

Data Base and Calibration Technique

Model runs were performed for three small experimental basins in southern Sweden: Gårdsjön F1 (0.04 km^2) and F3 (0.03 km^2) and Buskbäcken (1.8 km^2), see Fig. 2. The basins represent examples of the dominating type of landscape in Sweden, i.e. coniferous forested till soils on fractured gneiss or granite rock. The fractions of outcrop in the Gårdsjön F1 and F3 basins are relatively high, 47 and

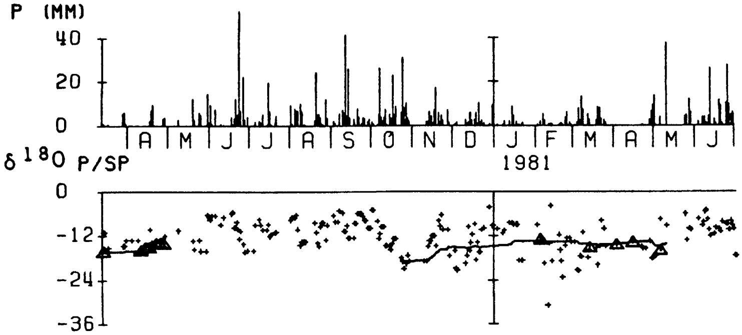


Fig. 3. Precipitation and $\delta^{18}O$ of the precipitation and the snowpack, Buskbäcken.
(Continuous curve ≡ simulated $\delta^{18}O$ of the snowpack,
triangles = $\delta^{18}O$ of the snowpack determined from snow samples).

32% respectively. A more detailed description of the hydrology of the Lake Gårdsjön area was given by Johansson and Nilsson (1985).

The ^{18}O-data consist of daily values of precipitation $\delta^{18}O$, a few values of snowpack $\delta^{18}O$ (see Fig. 3) and values of streamwater $\delta^{18}O$ at various time intervals.

Calibration was done by trial and error, starting with the water flow, mainly relying on visual inspection. When significant discrepancy between observed and calculated runoff occurred in the winter season, adjustments of temperature records were made, in order to avoid ^{18}O simulation being confused by errors in the water balance calculations. The precipitation amounts, however, were never altered.

The parameters affecting the ^{18}O-content in the snowpack (γ and ε) were estimated by comparison between computed and measured $\delta^{18}O$ in the snowpack (cf. Fig. 3) and ensuring that the computed and observed amounts of ^{18}O in the runoff did not differ noticeably.

Results and Discussion

The original structure of the PULSE-model could not describe streamwater $\delta^{18}O$ (Fig. 4, above). Sufficient damping was, however, obtained by considering mixing with soil water in the unsaturated zone and introducing additional storage volumes (Fig. 1), which do not affect runoff calculations (Fig. 4, below). Final simulations for Buskbäcken and Gårdsjön F1 are shown in Fig. 5.

Quite similar results were obtained for the two small basins in Gårdsjön. The best choice of parameter values was less distinct for the Gårdsjön basins than for Buskbäcken. The sensitivity to a change in the additional mixing volume in the unsaturated zone is illustrated in Fig. 5 (below).

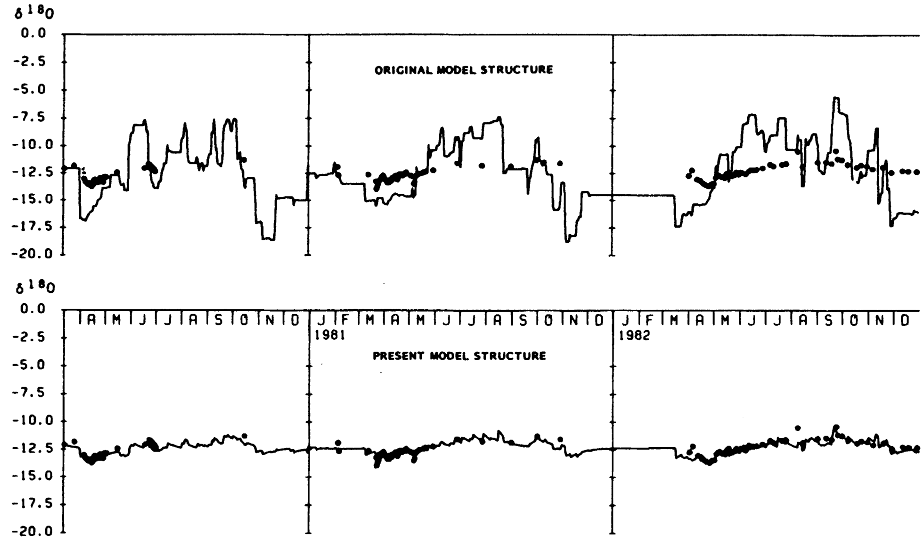
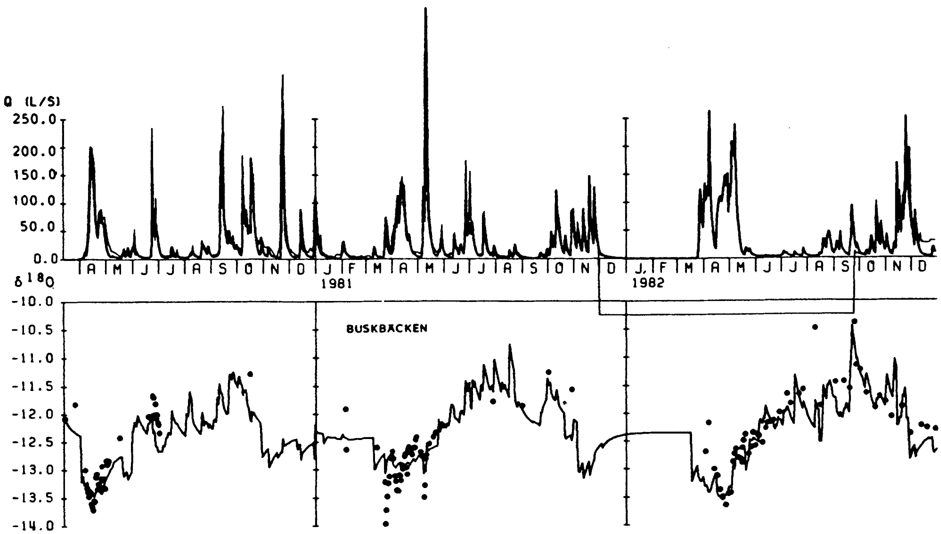


Fig. 4. Simulation of streamwater $\delta^{18}O$, Buskbäcken.
(Continuous curve = simulated values, Dots = observed values.)
Above: Original model structure. Below: Present model structure.

Estimates of the extent of saturated areas were obtained as a result of the calibration. The modelled fractions of saturated area varied considerably with time, from close to zero to maximum values of 17 % in Buskbäcken and 4 % in the two Gårdsjön basins respectively. The relatively small fractions of saturated areas, which contribute to overland flow, agree qualitatively with the results of Rodhe (1984 and 1985). He concluded that a major part of streamflow originated from groundwater for both the Gårdsjön basins and Buskbäcken. The dominating role of groundwater (or prestorm water) in streamflow generation has also been revealed by isotope studies in other countries (e.g. Dinçer *et al.* 1970, Sklash and Farvolden 1979).

Transit Time Distribution

The transit time distribution of the model can be computed by a simple experiment. Steady flow conditions are assumed, with constant precipitation equal to the mean runoff, and a spike of tracer is added to the model. The transit time distribution is then given as the outflow concentration, normalized to unit area (see Eriksson 1971). The result of such a procedure for Gårdsjön F1 is given in Fig. 6. In the experiment annual mean values of model storage volumes were assumed, and the evapotranspiration was set equal to zero.



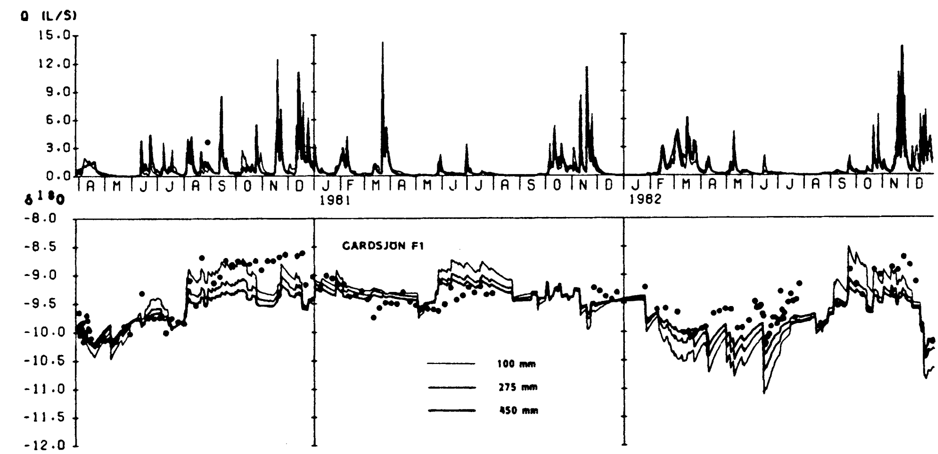


Fig. 5. Simulation of runoff (temperature updated) and $\delta^{18}O$. Calibration periods.
(Runoff: thick curve = simulated, thin curve = observed.
$\delta^{18}O$: continuous curve ≡ simulated values, dots ≡ observed values.)
Above: Buskbäcken,
Below: Gårdsjön F1.
For Gårdsjön F1 the simulated $\delta^{18}O$ for three different choices of additional mixing volume in the unsaturated zone (100, 275 and 400 mm) is shown.

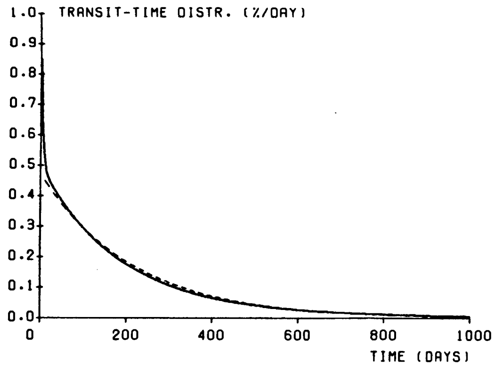


Fig. 6.
Approximation of the transit time distribution for the Gårdsjön F1 basin (continuos curve). The transit time distribution of an ideally mixed reservoir with the same turnover time is plotted (dashed curve) for reference.

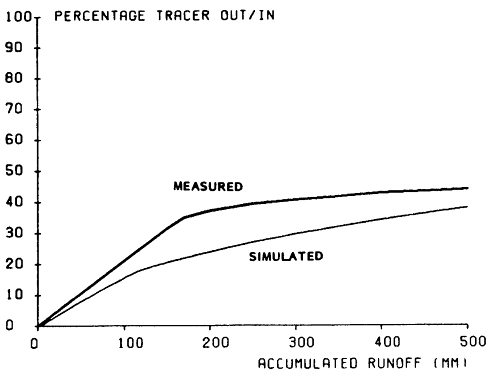


Fig. 7.
Cumulative amount of tracer in the runoff at the outlet of the Gårdsjön F1 basin. (Thin curve = simulated, thick curve ≡ measured by Nyström, 1985).

Nyström (1985) performed basinwide injections of tritium in the Gårdsjön F1 and F3 basins. In Fig. 7 a comparison is given between his investigation for Gårdsjön F1 and a simulation of the injection using the present model. Observed meteorological conditions were used as input to the model. Hence, in contrast to the situation in Fig. 6, steady flow conditions were not assumed. In the field investigation as well as in the model simulation, injections were made 40 cm below the ground surface and only on recharge areas.

The mean storage volumes obtained were 440 mm for Gårdsjön F1, and 480 mm for Buskbäcken, giving turnover times of 7 and 12 months respectively. However, the performance of the model was rather insensitive to variations in the size of the additional storage volumes. Hence, calculations of storage volumes and transit times are quite uncertain.

Conclusions

The PULSE model was able to describe a large part of the variation of $\delta^{18}O$ in streamwater after mixing with the soil moisture storage was considered.

The introduction of a bypass, representing saturation overland flow, increased model performance, especially for the Buskbäcken basin. This is reflected by the numerical values of the maximum fraction of satuated area, obtained by calibration of the model, giving 17% for Buskbäcken and 4% for the two Gårdsjön basins.

The storage volumes were estimated to approximately 400-500 mm, and turnover times to 7 and 12 months for Gårdsjön F1 and Buskbäcken respectively. These results agree reasonably well with tritium experiments performed by Nyström (1985) and investigations made by Rodhe (1985), both carried out in the Gårdsjön F1 basin.

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